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ORIGINAL STRUCTURES OF BEACHES, BARS, AND DUNES

BY WARREN O. THOMPSON

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INTRODUCTION

GENERAL STATEMENT

This paper presents, in detail, facts about original structures of modern beaches, bars, and dunes, and stresses especially their stratification. It attempts also to formulate criteria for recognition of ancient bars, dunes, and stratified beaches that lack organic remains, which might reveal the origin of the deposits.

For more than a century, the subject has been dealt with by numerous writers but little order has been brought out of the confusion. It is hoped that by very detailed studies of the formation of modern beaches, bars, and dunes a better means of interpreting these deposits may be worked out.

The writer spent nearly a year along the coast of California, between San Francisco and San Diego, examining various deposits, some of which are modern beaches, and others contain fossils indicative of ancient

beaches. In addition, beaches and bars of lakes, as well as eolian deposits, were examined in many localities in the western States, as far east as Colorado.

ACKNOWLEDGMENTS

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DEFINITIONS

The terminology of beaches is necessarily complex, because it deals with continually shifting zones and deposits. D. W. Johnson[1] summarizes much of previous usage of terminology of beaches, and his terms are employed insofar as possible. Johnson, however, is primarily concerned with description of zones of the coast, whereas this work deals with deposits in these zones. Therefore, some adjustments of terminology are necessary. Deposits of marine deltas and of estuarine tidal flats and lagoons are considered only incidentally in this paper.

GLOSSARY OF TERMS

Shore: the zone over which the water line, the line of contact between land and sea, migrates.

Coast: zone landward from the shore.

Beach: deposit that rests on the shore.

Coastline: boundary between the shore and the coast, marking the seaward limit of the permanently exposed coast.

Low-tide shoreline: seaward limit of the intermittently exposed shore.

Foreshore: zone between the crest of the beach and low-tide shoreline.

Lower foreshore: that section of the foreshore extending seaward from the landward limit of the zone of saturation to low-tide shoreline.

Lower foreshore beach: deposit that rests on the lower foreshore and is formed by a combination of waves and littoral currents. Sandy deposits below low-tide shoreline are not included in the lower foreshore beach.

Upper foreshore: that part of the foreshore extending seaward from the crest of the beach to the zone of permanent saturation.

Upper foreshore beach: the deposit that rests on the upper foreshore and is formed primarily by waves.

[1] D. W. Johnson: *Shore processes and shoreline development* (1919) chapter 4.

Backshore: that part of the shore lying between the foreshore and the coastline, covered by water only during exceptional storms or tides.

Backshore beach: deposit that rests on the backshore. It is formed of sand deposited by sheet flood from waves spilling over the crest of the beach at high tide, by storm waves, and by wind. In places, layers of stream-laid sediment and deposits of lagoons are also interlaminated.

Crest: upper edge of a marked seaward slope—in places, a ridge—formed by average high tides, but shifted occasionally by maximum high tides.

PRECEDENT WORK

No attempt is made to summarize the voluminous literature on the structure of beaches and related features, but a study of that literature has revealed many conflicting views. For example, some authors believe that laminae inclined consistently seaward at low angles are characteristic of beaches; others hold that laminae that dip at steep angles and in divergent directions are typical. This paradox undoubtedly arises from the fact that some studied the upper foreshore beach; others, the lower foreshore. A diversity of opinion exists, moreover, about the sorting of beach materials. Some contend that marked changes in size of particles of sand are to be found in beaches; others, that the sizes of particles differ only slightly.

ORIGINAL STRUCTURES OF MODERN BEACHES

LAMINATION IN BEACHES OF UPPER FORESHORE

Description of Laminae.—The most noticeable feature of the stratification of beach sands in California is the remarkable division into laminae, sorted both as to materials and size of particles. Plate 1, figure 1, shows a typical section of beach where this lamination is especially well developed.

Some laminae are dominated by light-colored minerals of moderately low specific gravity, such as quartz and feldspar, whereas others are dominated by dark-colored heavy minerals, such as magnetite and hornblende (Pl. 1, fig. 2).

The average thickness of laminae measured in this study is 0.3 inch (average of 502 observations). Of the total thicknesses, 80 per cent are between 0.1 inch and 1.0 inch. Some layers of sand are not laminated for a foot or more; others are single laminae no thicker than single grains of the fine sand that compose them.

Light-colored laminae average 0.44 inch in thickness; dark laminae, 0.06 inch. The thickest single dark lamina measured is 1.2 inches. In terrace deposits east of Monterey Bay, some lenses of magnetite sand are from 8 to 10 inches thick.

In most places, light-colored laminae dominate the higher part of the upper foreshore beach, where, in the absence of intervening dark laminae, they are distinguished by variations in size of particles from one lamina

FIGURE 1. LAMINATION
Upper foreshore beach, east shore of Monterey Bay, California.

FIGURE 2. LIGHT AND DARK LAMINAE
Monterey Bay, California.

FIGURE 1. TYPE C
Laminae of low seaward dip, buried by laminae of steep seaward dip. Zone seaward from place marked X and above gravel layer illustrates type C. Blurred zone in foreground is flood water. Mouth of Soledad Canyon, La Jolla quadrangle, California.

FIGURE 2. TYPE B
Laminae of seaward dip buried by laminae of landward dip. Sea to the left. (Scarp 8 feet high.) West Jetty, Balboa, California.

FIGURE 3. TYPE D
Laminae of steep seaward dip (gravel layer) buried by laminae of low seaward dip. Swirl structures shown above (and left of) scale. Mouth of Soledad Canyon, La Jolla quadrangle, California.

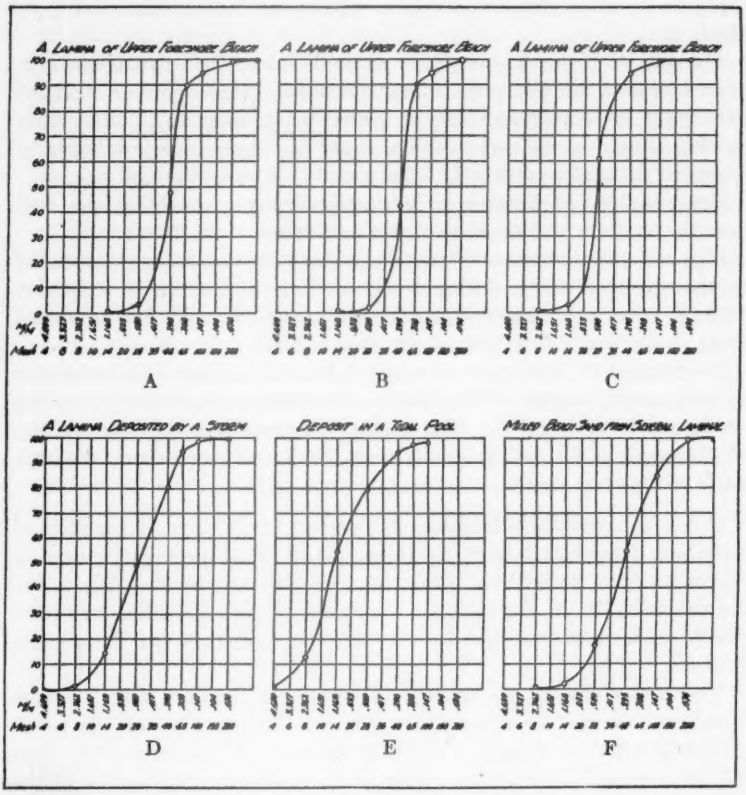

FIGURE 1.—*Cumulative logarithmic diagrams of screen analyses of beach sands*

to another. These light-colored laminae are generally thicker than those in the lower part of the upper foreshore beach, where dark-colored laminae usually equal in number, but not in thickness, the light-colored laminae.

In cross-sections of the beach parallel to the shoreline, a typical lamina may be traced 100 feet or more before it disappears. In sections normal to the shoreline, a single lamina seldom may be traced more than 25 feet before it is terminated by a surface on which rest other similar laminae with a slightly different dip.

Particles of a single lamina fall characteristically within a narrow size range, although they may differ considerably in size from equally well-sorted particles of adjacent laminae.

Figure 1 shows graphically these significant features in sorting of beach sand.

A and B of Figure 1—diagrams of sands taken from two laminae in a cross-section of the same upper foreshore beach—show a marked uniformity in sorting and size of grain. C, a diagram of sand from another lamina in the same section, shows the characteristic uniformity in sorting but differs distinctly from A and B in size of grains.

Cross-sections of beaches include some coarse-grained laminae that lack the uniform sorting characteristic of most of the beach.

D is a diagram of a coarse-grained, imperfectly sorted layer, deposited on the upper foreshore during a storm. E is a diagram of a coarse-grained, imperfectly sorted layer, deposited by rills, which flow into tidal pools on the lower foreshore of the same beach.

F—diagram of a mixture of sands, taken at random from a number of well-sorted laminae of this beach—shows the same imperfect sorting as the sands of D and E. This analogy suggests that imperfect sorting of the sands of D and E results from mixing of sands from different parts of the same beach.

Origin of Laminae.—Laminae in beaches result from a complex of variables, dependent largely on variations in the transporting power of waves carrying particles of different sizes and specific gravities. The amount of sediment available for the waves to pick up, the slope of the beach, and, therefore, the velocity of the backwash are influencing factors.

The writer studied the growth of successive laminae near Newport Beach, California, for almost four months. Detritus left on the beach ranges in size from cobbles, 2 inches in diameter, to very fine sand. Following a period of unusually vigorous wave activity, the surface of the beach is strewn with cobbles, pebbles, coarse sand, and fragments of shells. Periods of lesser wave activity produce finer sediment. Successive changes in grain size of materials observed on the surface of the beach correspond to the similar successive changes in grain size of the constituents in the successive laminae that make up the beach.

Material either of diverse sizes or of diverse densities, or both, is necessary for the formation of a laminated beach. The rugged topography and complex composition of rocks of the California coast yield sediment of this sort.

Waves and littoral currents of the sea are capable of the thorough sorting and uniform spreading of sediment in individual laminae. Also, waves and littoral currents vary so greatly in carrying power that they can produce contiguous laminae of well-sorted material but of markedly different size range.

The observed uniform sorting of particles within a single lamina is the result of a finely balanced selecting process. An incoming wave brings to the slope of the beach all particles whose size and specific gravity permit their being transported by that wave. The backwash of the same wave returns most of the material to the sea, but leaves on the beach those grains that cannot be returned to the sea by the backwash, whose transporting power is less than that of the incoming wave. Laminae are added to the upper foreshore as long as deposition by the swash of a certain wave exceeds erosion by the backwash of the same wave. However, the backwash tends to erode deposits of the swash more deeply at the lower than at the upper surface of the beach, so that the slope of the upper foreshore, and hence the erosive power of the backwash, is gradually increased. When deposition by the swash is just balanced by erosion of the backwash of a wave, the slope of the beach is said to be at the "profile of equilibrium."[2] The carrying power of waves is so variable from day to day that this condition of equilibrium seldom prevails for long. Changes in height of water, due to tides and offshore or onshore winds, and changes in wave size, all enforce a continual modification of the profile of equilibrium and, hence, a corresponding change in size of grains that remain on the beach.

Dark-colored laminae result from the fact that their dark constituent grains of high specific gravity are more difficult to transport than are those grains of lesser density that make up the light-colored laminae. When the transporting power of the swash of a wave carrying a load of both light and heavy grains decreases up the beach, heavy grains tend to settle to the bottom of the lamina deposited by that wave. In returning over the surface of the newly made lamina, the backwash of the same wave picks up the most easily transported grains, but leaves a thin residual film of heavy grains on the surface of this new lamina. The next wave, in the same way, deposits heavy grains on this residual film. Hence, a dark lamina builds up by the addition of the residual film of heavy grains left by the backwash of one wave and those heavy grains that settle from the load of the swash of the next succeeding wave.

Time Represented by Laminae.—In measuring the interval of time represented by laminae, the writer assumes that a comparison between thickness of daily accumulation of particles on the beach and thickness of laminae within the beach may suggest the amount of time represented by a single lamina.

For the purpose of measuring daily accumulation of sand on the beach, a line of ¾-inch iron rods, each 4 feet high, was set normal to the shoreline at intervals of 15 to 30 feet, depending on the profile of

[2] D. W. Johnson: *Shore processes and shoreline development* (1919) p. 217.

the beach. The uppermost stake was placed at the upper reach of waves at high tide; the lowest, near the shoreline of minimum low tide. The change in height of the surface of the beach with respect to the rods was recorded daily for a week. Two such investigations were made—one at Rio del Mar Beach, on the east side of Monterey Bay; the other at Newport Beach, on the southern coast (Tables 1 and 2).

TABLE 1.—*Daily loss or accumulation of sand on surface of beach at Rio del Mar, Monterey Bay, California, January 9 to 16, 1932*

Pt.	Jan. 9	Jan. 10	Jan. 11	Jan. 15	Jan. 16	Total Gain	Total Loss	Av. Gain	Net Gain	Net Loss
1.....	1.8	1.4	2.5	−11.3	−3.5	5.7	14.9	1.9	...	9.3
2.....	1.6	2.6	1.3	−10.9	−2.2	5.8	13.1	1.9	...	7.3
3.....	2.2	1.6	3.4	−10.3	0.1	7.2	10.4	1.8	...	3.1
4.....	4.6	2.0	0.9	−4.0	1.3	8.8	4.4	2.3	4.8	...
5.....	2.4	−1.1	0.5	0.3	−0.1	3.2	1.3	0.6	2.0	...
6.....	−0.2	0.3	0.8	1.8	−0.9	2.9	1.1	0.6	1.9	...
7.....	−1.8	2.1	0.7	2.3	−0.8	5.1	2.6	1.0	2.5	...
8.....	−0.8	1.4	0.5	....	1.6	3.4	0.8	0.5	2.6	...
9.....	−0.8	−1.1	1.6	....	0.3	1.9	2.0	0.3	...	0.1
10.....	−0.7	−2.5	2.1	....	0.3	2.3	3.1	0.4	...	0.8
11.....	0.6	−0.6	0.7	....	−3.3	0.4	3.9	0.7	...	2.5
12.....	1.9	−2.1	0.1	....	−6.7	2.1	8.8	1.1	...	6.7
13.....	1.9	−1.1	0.5	....	−9.3	2.3	10.4	1.1	...	8.1
14.....	1.0	−0.3	0.1	....	2.6	3.6	0.4	0.6	3.2	...
15.....	−1.1	1.4	−1.5	....	3.3	4.6	2.6	0.7	2.0	...

Fixed points 1-15 are iron stakes driven into beach. Point 1 is near high-tide shoreline, point 15 near low-tide shoreline for month of January, 1932. Points are irregularly spaced in a line at right angles to shoreline. Changes in height of surface of sand from fixed point are recorded as gain or loss in inches. Data for January 15 are result of activity of 3 days. A heavy storm prevented observations on January 12, 13, and 14.

TABLE 2.—*Daily loss or accumulation of sand on surface of beach at Newport Beach, Orange County, California, February 17 to 24, 1932*

Pt.	Feb. 17	Feb. 18	Feb. 19	Feb. 20	Feb. 21	Feb. 22	Feb. 23	Feb. 24	Total Gain	Total Loss	Av. Gain	Net Gain	Net Loss
1.....	6.1	−8.5	3.6	5.1	2.0	0.7	0.1	0.0	17.7	8.5	2.9	9.2	...
2.....	−0.8	−1.6	2.5	0.6	3.2	2.2	1.0	0.0	9.6	2.5	1.9	7.1	...
3.....	4.7	−5.0	2.7	4.2	0.1	0.2	3.1	1.2	16.4	5.0	2.3	11.5	...
4.....	5.1	−6.7	−2.5	4.7	−1.4	−0.4	1.9	0.7	12.5	10.9	3.1	1.5	...
5.....	4.2	−8.6	−0.7	3.8	−1.6	0.5	−1.6	6.9	15.4	12.6	3.8	2.8	...
6.....	4.0	−7.6	−2.5	7.0	−0.7	0.1	0.4	5.0	16.5	10.9	3.3	5.6	...
7.....	−4.0	−5.5	3.6	4.7	−2.1	1.2	3.5	0.4	13.5	11.6	2.7	1.8	...
8.....	−3.1	−0.4	0.5	1.5	−1.7	2.7	−0.2	3.4	8.1	5.5	2.1	2.6	...
9.....	−3.7	0.4	1.5	−0.7	−0.6	0.9	0.1	...	2.9	5.1	0.7	...	2.2
10.....	−3.6	2.7	0.9	−1.1	0.8	0.4	0.0	...	4.7	4.7	1.1	...	0.0
11.....	−4.9	3.4	0.4	−0.4	0.6	0.9	−0.1	...	5.1	5.3	1.3	...	0.2
12.....	−6.7	1.8	−0.1	0.7	−0.2	0.5	0.2	...	3.2	7.1	0.8	...	3.9
13.....	−6.0	1.5	−1.2	−0.1	−0.1	0.0	0.2	...	1.7	7.5	0.8	...	5.7
14.....	−5.9	1.1	−2.5	0.0	0.2	−0.5	0.5	...	1.9	8.9	0.9	...	7.0
15.....	−7.9	0.7	−2.9	1.5	1.0	−1.5	−0.4	...	3.2	12.6	1.1	...	8.4

Fixed points 1-15 are iron stakes driven into beach. Point 1 is near high-tide shoreline, point 15 near low-tide shore line for month of February, 1932. Points are spaced 15 feet apart in a line at right angles to shoreline, except for a distance of 30 feet between points 14 and 15. Changes in height of surface of sand from fixed point are recorded as gain or loss in inches.

The data show, quantitatively, that accumulation is, usually, slightly greater on the upper foreshore than on the lower, and that deposition on one part of the beach profile is compensated by erosion on another. In the observations at Newport, the net deposition against the upper eight stakes was balanced by erosion around the lower seven stakes. Stake 8 was at the contact between upper and lower foreshores. The average daily accumulation along the stake line on the beach at Newport Beach was 1.93 inches; at Rio del Mar, it was 1.02 inches. These accumulations, however, are subject to concurrent erosion. The average thickness of these laminated accumulations is nearly five times the average thickness of a single lamina.

Where the thickness of daily accumulation of beach material against stakes was known for several consecutive days, pits were dug to expose a vertical section. The laminae constituting a daily accumulation were then counted. The accumulation of a single day, or of a number of days, in some places consisted of a single lamina; in other places, of several laminae. No definite relation was found between the number of laminae and the thickness of daily accumulation. Lamination is apparently the result of deposition by an unknown number of waves, whose carrying power varies from time to time, and with each change begins the deposition of a new lamina. The carrying power of waves that wash the beach is modified by rise and fall of the tide, but the effect is not enough to obliterate the stronger effect of variation in the carrying power of waves from the open sea. As a result, rhythmic banding due to tidal variations alone is not noticeable in the beach. Even if formed, tidal lamination is probably destroyed by the strong backwash of occasional destructive waves, before it becomes a permanent part of the beach.

It is concluded that the interval of time represented by a group of beach laminae indicates no exact interval of time, because such a group includes intervals of non-deposition and of erosion. Furthermore, a group of beach laminae is the result of deposition of irregularly recurrent variations in carrying power of waves. A few laminae may represent the deposit of a single day. But even if each lamination represented one day's deposition, the successive laminations in a thick section would not necessarily represent successive days, but perhaps only those occasional days whose contributions escaped erosion.

CROSS-LAMINATION IN BEACHES OF UPPER FORESHORE

Uncusped Beaches.—Four types of cross-lamination, shown diagrammatically in transverse vertical section in Figure 2, are found in California beaches. Each diagram is drawn for the structure as viewed in

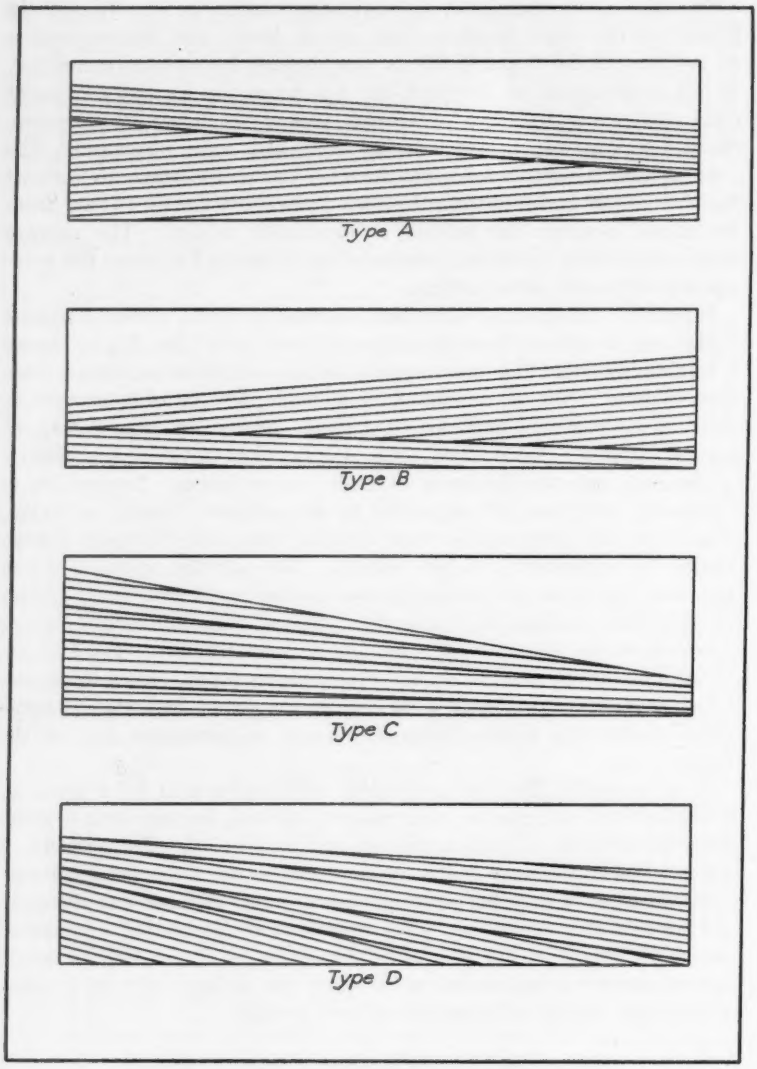

FIGURE 2.—*Diagrams of cross-lamination*
Sea to right in each figure.

cross-section at right angles to the shoreline, with the sea to the right. The types are designated A, B, C, and D, respectively.

In type A, laminae of landward dip are truncated by an erosional surface of seaward dip; this surface, in turn, is the floor of deposition for laminae of seaward dip. The angle of seaward dip of the laminae is slightly less than that of the floor of deposition, so that each lamina thins landward on this surface—an overlap of small dimensions. The strikes of both laminae and truncating surfaces are approximately parallel; hence, the structure, when viewed in cross-section parallel with the strike of the laminae, appears as a group of parallel laminae.

Type B, which is somewhat less common than type A, has no intervening surface of erosion. Laminae with a seaward dip are buried by laminae with a landward dip (Pl. 2, fig. 2).

Types C and D have only laminae with a seaward dip. In type C the surface truncating the underlying laminae is also the floor of deposition for the overlying laminae; this surface dips seaward more steeply than do the truncated laminae. In type D the truncating surface dips seaward less steeply than do the truncated laminae. As in types A and B, these truncating surfaces and laminae also have essentially the same strike (Pl. 2, figs. 1 and 3).

The extent of these cross-laminated zones varies. Some are a few feet long and a few inches thick; others are 150 to 200 feet long and 8 to 10 feet thick. An average cross-laminated zone of a California beach is about 50 to 75 feet long and 4 to 6 feet thick.

Cross-lamination in beaches results from changes in the beach profile of equilibrium on shore. Fluctuations in height of sea water, induced by spring and neap tides, variations in transporting power of waves and currents, in texture and quantity of sediment, in the slope on which sediment is moved, and variations in coincidence of activity among these factors—all combine to change the beach profile of equilibrium and, hence, to produce cross-lamination.

An exact correlation between observed cross-lamination and the processes by which it is formed requires many more controlled observations than were permitted in the scope of this investigation. However, the general process is outlined in a study of changes in form of the beach profile at Newport Beach, California. Beach profiles were measured carefully in the same place, from a point of reference on the coast to low-tide shoreline. Observations were taken each week from April 18 to June 11, 1932. It was assumed that laminae, when deposited, conform approximately in dip and strike to the slope of the surface of the beach. This supposition was verified later by trenching into the beach along the line where profiles were measured. Results of the work are shown in Figure 3. The profile of April 18 ends landward against a

low, wave-cut scarp, beyond which the surface of the beach had been graded artificially.

On April 22, a terrific storm, accompanied by powerful wave activity, eroded the surface of the beach to a gentle seaward slope and left a wave-cut scarp at the landward end of the profile of April 25 (Pl. 3, fig. 1). Offshore, the sea was discolored by suspended sediment, which gradually returned to the beach as the storm subsided.

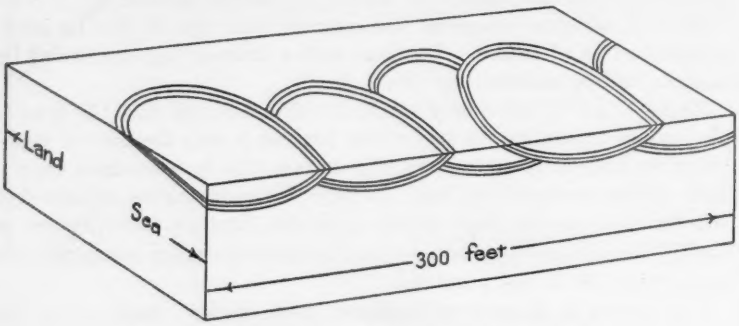

FIGURE 3.—*Weekly changes in profile of foreshore beach*
April 18 to June 11, 1932, Newport Beach, Orange County, California.

From May 2 to May 9, neap tides dominated the sea, so that waves at high tide reached only about half the distance to the wave-cut scarp. These conditions resulted in a low, wave-built ridge on the beach, about half-way between low-tide shoreline and the wave-cut scarp of the storm.

From May 9 to June 3, during which period spring tides dominated the sea, the low ridge, or crest of the beach, gradually was washed landward about 50 feet and was built up nearly 4 feet by the waves. Waves washed material from the seaward slope and carried it over the crest of the low ridge to form laminae with a gentle landward dip on the landward slope of the ridge.

From June 3 to June 11, waves once more left laminae on the seaward slope of the beach. This depositional phase, which seems to have resulted from a recurrence of neap tides, greatly reduced the seaward slope of the foreshore, which resulted from landward migration of the crest of the ridge.

On June 11, a trench, dug across the crest of the beach, exposed type A cross-lamination (Fig. 2, A). Whether or not type B cross-lamination was formed below the floor of the trench, by contact of landward-dipping laminae of type A with underlying laminae, was not determined at the site of the measured profiles. It is probable, however,

FIGURE 1. EMBAYMENT EXPOSED IN WAVE-CUT SCARP
Embayment near center of view. Scarp 6 feet high. Near Balboa, California.

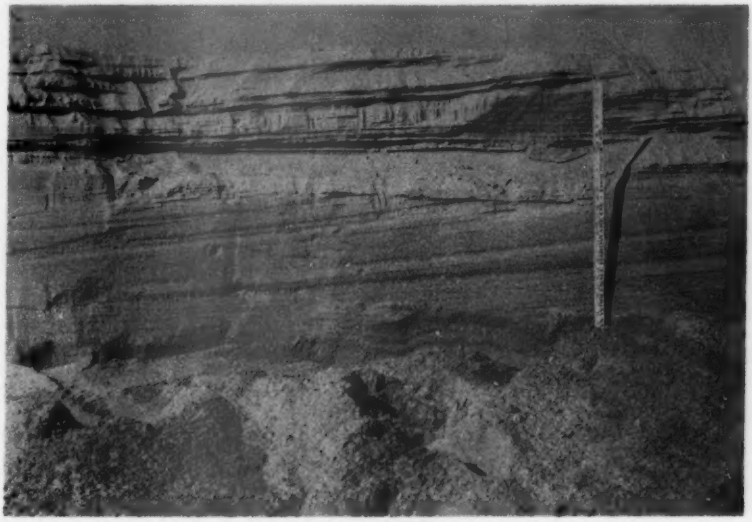

FIGURE 2. CROSS-LAMINATION FROM OVERLAPPING SCOOPLIKE EMBAYMENTS
Near Balboa, California.

SCOOPLIKE EMBAYMENTS

FIGURE 1. CROSS-SECTION OF BURIED SCARP
Extreme example of type C cross-lamination. Sea to left. Pits below scale result from removal of pebbles. Mouth of San Dieguito Creek, La Jolla quadrangle, California.

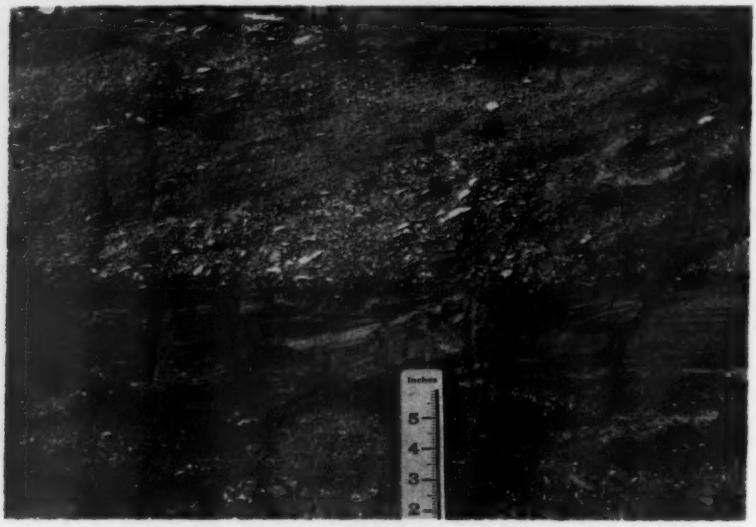

FIGURE 2. INTERSTRATIFICATION OF FOSSILIFEROUS AND MICACEOUS SANDS
Upper unit of terrace deposits at Newport Beach, California.

CROSS-SECTION DETAILS

that the structure of the beach below the trench is similar to that at West Jetty, near Balboa, California, where a like succession of phenomena took place. Plate 2, figure 2, shows cross-lamination of type B at West Jetty.

Inasmuch as shore deposits are subject to the erosion of storms, the order of succession of landward- over seaward-, or seaward- over landward-dipping laminae, as shown in types A and B, depends on the profile and the structure of the beach when a cycle of deposition begins, and on whether neap or spring tides prevail during deposition.

Cusped Beaches.—Cusped beaches reveal a more complex structure than do uncusped ones, because the slopes of the cusps and their intervening embayments form numerous, more or less regular variations in the strike of the slope on which the cross-laminated structures develop, in contrast to the approximately plane surface of uncusped beaches.

Between the pier at Balboa, California, and the jetty at the entrance to Newport Bay (West Jetty), there is a stretch of beach where cusps form with remarkable regularity. Occasional storms, which destroy these cusps, leave the beach a smooth surface sloping gently seaward. As soon as the sea regains its normal condition, however, cusps again develop. Each cusp is separated from adjacent cusps by broad scooplike embayments, whose axes plunge gently toward the sea. These embayments average 140 feet in width from the crest of one cusp to that of the next. Waves gradually fill the embayments with laminated sand. Each lamina dips in conformity with the floor of the embayment and thins toward the landward perimeter.

Changes in conditions of equilibrium on shore, especially during storms, erode upper parts of cusp ridges and smooth the beach to an uncusped seaward-sloping surface. The position of axes of cusps and embayments that develop subsequently usually does not conform to that of the former ones; hence, laminae of new embayments rest unconformably on laminae of remnants of old embayments. This condition results in a cross-laminated structure, seen in cross-section parallel to the shoreline in Figure 4.

Numerous examples of this type of cross-lamination were seen in the wave-cut scarp of the cusped segment of beach between the pier at Balboa and West Jetty (Pl. 3). The dimensions of the structures and the attitudes of involved laminae correspond to those of cusps in the active beach in front of the wave-cut scarp.

OTHER ORIGINAL STRUCTURES IN BEACHES OF UPPER FORESHORE

Buried Scarps.—In some places the small seaward-facing scarp, cut into the beach by the activity of waves and ocean currents during storms, becomes buried by deposition of waves. Plate 4, figure 1, shows this

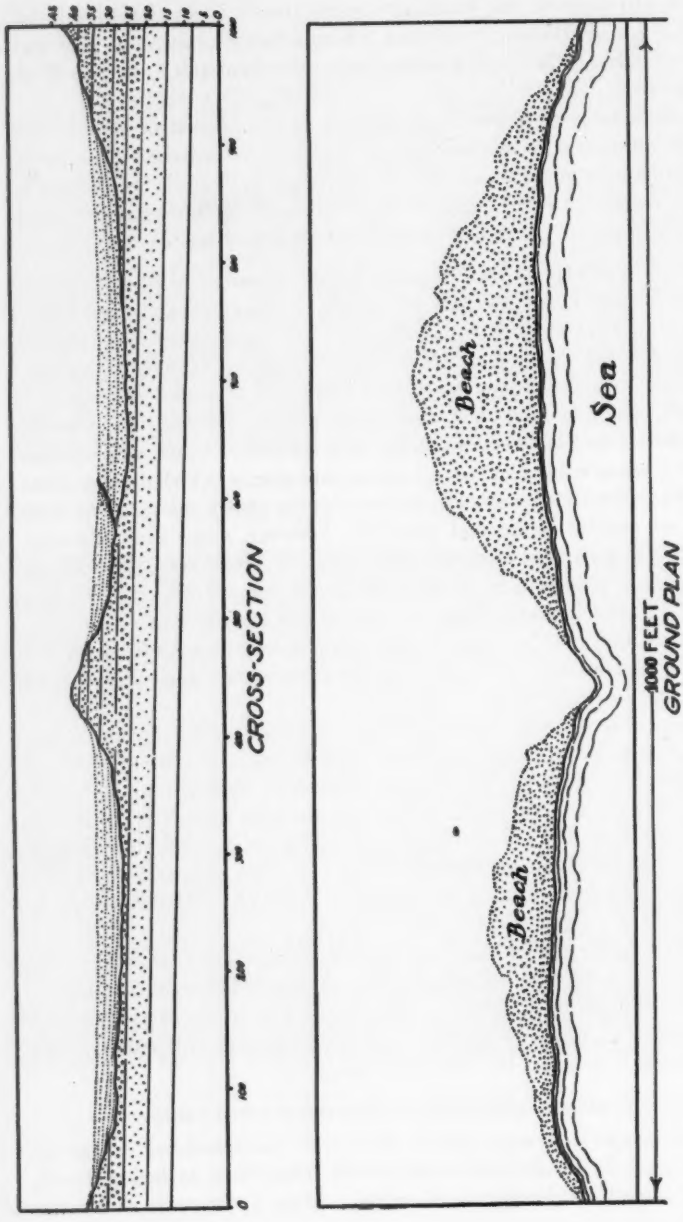

FIGURE 4.—*Diagram of cross-lamination from overlapping of scooplike embayments*

feature in cross-section. The cross-laminated structure, insofar as the laminae above the truncating surface are more steeply inclined and more wedge-shaped than is normal, is an extreme example of type C. The abnormal steepness of the profile of equilibrium, on which this structure is built, makes available an excessive amount of material for deposition of laminae, partly because the forward motion of the swash is checked abruptly against the steep face of the scarp and partly because an abundance of sand is dashed loose by the impact of the swash.

This relationship between the cross-laminae and a buried scarp is distinctive of beaches, in that the general trend of the scarp is approximately parallel to the strikes of laminae that bury it and of laminae on which it is cut. The trend of scarps cut in sand by wind and streams is approximately parallel to the dip of burying laminae, except that, in a few places, sand blown over the tops of scarps may bury them with crudely laminated sand that strikes parallel to the face of the scarp.

Buried Channels.—In some places along the beach, wave-cut scarps expose cross-sections of troughs with rounded floors, which have been filled with coarse, poorly sorted, and crudely laminated sand. The axes of these buried channels apparently extend at right angles to the shoreline, but it is impractical to determine whether the channels slope toward or away from the sea. At the place where the structures were seen, it is possible that they may have resulted either from the seaward flow of bay waters or from the landward swash of sea water. The structure resembles that which is commonly observed in fluvial deposits, except that the floor of the channels is more rounded.

Unlaminated Interjacent Lenses of Sand.—Interfingering lenses of unlaminated beach sand comprise another structure, which is exposed in wave-cut scarps. The floor of one lens is cut into some part of an adjacent or an underlying lens.

Probably, interjacent lenses of sand and buried channels were formed by a similar process, except that in the formation of buried channels the load of detritus in proportion to the carrying power of the water was smaller than in the formation of interjacent lenses of sand.

Backwash Marks.—Backwash marks are sand patterns, which resemble the pattern of scales on some ganoid fishes, developed on the surface of the beach. Johnson [3] describes this structure under the heading "backwash marks." E. M. Kindle [4] illustrates the structure under the heading "imbricate wave sculpture." Backwash marks appear near the lower limit of the upper foreshore beach in many places in California. They

[3] D. W. Johnson: *Shore processes and shoreline development* (1919) p. 517.
[4] E. M. Kindle: *Recent and fossil ripple mark*, Geol. Surv. Canada, Mus. Bull. 25 (1917) pl. 19 B.

apparently are formed by the combined erosion and deposition of cross-currents of sheets of water returning down the surface of the beach. This structure should be good proof of wave activity in ancient deposits.

Swirl Structures.—Near the contact of coarse sand with overlying fine sand, involved masses of light- and dark-colored sand are found in numerous places in sections of modern beaches in California (Pl. 2, fig. 3). Inasmuch as swirl structures are common in modern beaches, their presence in a sedimentary rock may support other evidence for an ancient beach.

Ripple Marks.—Ripple marks appear in many places on beaches of California, but, as this subject has been exhaustively treated by other authors, no discussion is attempted here. With careful study, ripple marks should be valuable as supporting evidence.

CROSS-LAMINATION IN BEACHES OF LOWER FORESHORE

General Statement.—The lower foreshore beach is defined as that part of the beach that is permanently saturated with water. It occupies the zone along the shore between the seaward edge of the upper foreshore and the minimum low-tide shoreline.

The structures of lower foreshore beaches cannot be observed in cross-section, because the saturation of the sand causes an immediate cave-in following excavation. Inferences as to the structures of lower foreshore beaches are therefore drawn from a knowledge of their surface forms and of the processes that produce them.

Description of Lower Foreshore Beaches.—Material of lower foreshore beaches is less uniformly sorted than that of most upper foreshore beaches, and its composition is distinctive. A considerable part of the detritus in the tidal pools consists of minute flakes of mica and fragments of sea shells—materials more readily transported than those of the upper foreshore. Not infrequently, laminae of micaceous sand on the floors of tidal pools are buried by steeply inclined laminae of coarse sand containing many fragments of shells.

Commonly, the most prominent topographic feature of lower foreshore beaches is a broad, low ridge, separated from the seaward slope of the upper foreshore beach by a wide, shallow trough, whose axis is parallel to the shoreline (Pl. 5, fig. 1). Part of the time, however, this trough is filled with beach sediment, so that a continuous slope is formed from the upper foreshore to the lowest limit of shore exposure. The declivity of this slope decreases seaward.

Numerous small basins and shallow embayments interrupt the continuity of the ridge and trough of the lower foreshore. Irregularly shaped,

FIGURE 1. LOW RIDGE AND TROUGH ON LOWER FORESHORE
Shallow tidal ponds and water courses exposed at low tide. Newport Beach, California.

FIGURE 2. TIDAL POOLS AND SAND BARS OF LOWER FORESHORE
A small delta is being built in foreground.

LOWER FORESHORE

FIGURE 1. FLOOR OF DEPOSITION OF ANCIENT BEACH
Dark line indicates position of floor. Terrace deposit near Newport Beach, California.

FIGURE 2. INTERJACENT WEDGES OF SAND SEPARATED BY TRANSVERSE SURFACES OF EROSION
Terrace deposits near Newport Beach, California.

plateau-like sand bars, with lobate, steep sides, project into these depressions (Pl. 5, fig. 2). Some depressions have a surface area of a few square feet; others, hundreds of square feet. The depth in the central part of the depression is ordinarily 2 to 3 feet.

The currents that wash the lower foreshore are extremely complex. Added to the swash and backwash of normal wave activity are the littoral currents resulting from tides and from waves that strike the shore at an angle.

Furthermore, seepage water from the upper foreshore forms rills, which transport beach sediment to tidal pools. In many places, these minute drainage systems build small deltas in the tidal pools.

The various currents, which wash the lower foreshore during tidal advance and retreat, wash sand from the top of the flat-topped sand bars and add steeply inclined laminae to their sides. Waves swashing across depressions adjust sediment to the depression floors so that the least possible resistance is offered to wave currents.

Inferences as to Probable Structures of Lower Foreshore Beaches.—It is inferred from the conditions of deposition described that: (1) deposits on the lower foreshore are intricately cross-laminated; (2) discontinuous, nearly horizontal layers of short, sandy, foreset laminae are interstratified with evenly laminated layers of fine micaceous sand, which conform to irregularly shaped depressions; (3) the short foreset laminae dip in all directions with an angle of dip that approaches 30 degrees; and (4) there is no correspondence in direction of dip between the nearly horizontal micaceous laminae and the foreset laminae. The size of these structures is much smaller than that of the structures of the upper foreshore beach.

Confirmation of the inferred structures of the lower foreshore beaches appears in terrace deposits near Newport Beach, California, where a stratum, 3 feet thick, corresponds closely in texture, mineral composition, and distribution of materials to the anticipated qualities and structures of lower foreshore beaches (Pl. 4, fig. 2). Moreover, the fossils and the field relations of the stratum make it extremely probable that it is an ancient lower foreshore beach. A more complete discussion of ancient beaches will be presented.

COMPILATION OF INCLINATION OF LAMINAE IN MODERN BEACHES

Approximately 1200 measurements of slopes of beach surfaces and of beach laminae corresponding to these slopes were recorded during this investigation. Of this number, nearly 1000 were derived from upper foreshore beaches. The angles of slope are as follows: 48 per cent are between 0 and 3 degrees, 34 per cent between 4 and 7 degrees, 11 per cent between

8 and 11 degrees, 4 per cent between 12 and 15 degrees, and only 3 per cent 16 degrees or over.

Approximately 200 measurements of the angles of foreset slopes in deltas and sand bars of tidal pools were recorded. No attempt was made to discriminate between the foreset slopes of deltas and those of advancing sand bars. The foreset slopes have an average maximum declivity of 28 degrees. In lower foreshores, 57 per cent of the laminae are 16 degrees or over, as compared with 3 per cent of the laminae in upper foreshores.

FOSSILS IN BEACHES

The shell forms buried in the beach, even though predominantly beach-dwelling, are characteristically mixed with forms from other environments.

Certain animals are indigenous to the beach, but it is probably impossible to find a modern beach that does not contain many shell forms from other sites. Shells are washed from rocky headlands, from deep water, and, occasionally, from rivers and estuaries.

Any sandstone or conglomerate containing an appreciable number of shells from various ecological environments is likely to be littoral, and it may be an ancient beach, inasmuch as such deposits are usually made relatively close to the shoreline.

ORIGINAL STRUCTURES OF ANCIENT BEACHES

GENERAL REMARKS

Opportunity for examination of the stratification of modern beaches in cross-section is confined to the relatively narrow, intertidal, wave-laid fringe of the upper foreshore beach. Stratification of the lower foreshore beach, occasionally exposed at extremely low tide, must be inferred from detailed knowledge of the "microphysiography" of deposits of that zone. Beyond the shoreline of minimum low tide, in deeper water, are other sandy deposits, probably formed by processes similar to those prevailing on the lower foreshore. As these deposits are always covered with water, direct observation of their internal structure in cross-section is not feasible.

Some of the sandy terrace deposits that have recently emerged from the sea along the coast of California afford well-exposed cross-sections of stratified materials, which, if satisfactorily shown to be beaches, increase the opportunities for examination of structures of beaches and related formations.

Two localities in California were chosen for this investigation: the marine terrace deposits of Pleistocene age at Newport Beach, Orange County, and those at San Pedro, Los Angeles County.

A part of the terrace deposits at each of these places has characteristics suggestive of ancient beaches. Each is made up largely of sandy material containing, or interstratified with, a mixed marine fauna, dominated by forms living on or near the beach;[5] each is distributed along the present shoreline, a few feet above sea level, and is a part of a marine terrace that has recently emerged from the sea.

TERRACE DEPOSITS AT NEWPORT BEACH

General Statement.—The terrace deposits at Newport Beach are composed of two distinct units: a lower, the characteristics of which are analogous to those of modern beaches of the upper foreshore; and an upper, the characteristics of which correspond to those that are inferred from processes of deposition in modern beaches of the lower foreshore.

Both units occupy a narrow zone, which is roughly parallel to the present shoreline and not more than 40 feet above present mean sea level. As the formation is traced inland from the present shoreline, the materials change from fossiliferous marine sands to non-fossiliferous marine sands and then to non-fossiliferous sandy gravel, whose stratification resembles that formed by streams. Physiographers agree that terraces with flat tops like the terrace at Newport are segments of the former shore, which have emerged from the sea.

Characteristics and Origin of Lower Unit.—The lower unit is deposited on a surface of erosion, a cross-section of which corresponds to a cross-section along a shoreline with numerous indentations (Pl. 6, fig. 1). Manifold broad, shallow, crescent-shaped depressions in the floor of deposition correspond to indentations of a shoreline, and the elevations between these depressions correspond to headlands (Fig. 5). If these depressions were stream-cut, they would probably be V-shaped channels or troughs with flat floors and abrupt slopes, and would contain, normally, some detritus from the hinterland. Because the depressions lack the characteristic cross-sections of stream-cut channels, because they contain only sand that is mixed with marine fossils and masses of yellow clay derived from the underlying formation, and because they appear with greater frequency than do stream-cut channels along the present shoreline, the balance of evidence supports the conception of a wave-cut, rather than a stream-cut, surface for the floor of deposition of the lower unit.

Interjacent wedges of sand, separated by transverse surfaces of erosion, comprise the lower unit. Each wedge contains one or several layers of laminae of sorted sand having a distinct primary dip (Pl. 6, fig. 2). In

[5] For a more complete discussion of the fossils, see writer's unpublished thesis for Ph.D. degree, Stanford Univ., Calif. (1935) p. 103-109.

length, thickness, and attitude, these laminae are like those of modern upper foreshore beaches of California. As in modern beaches, horizontal laminae are uncommon.

Characteristics and Origin of Upper Unit.—Materials of the upper unit vary greatly. The average size of particles is smaller, but the range in size of particles is greater, than in the lower unit. Coarse, fossiliferous, granitic sand and micaceous, sandy silt are the most common materials. A few large fragments of sandstone and granite were also found. The upper unit is, therefore, like that of the modern lower foreshore beach in quality of minerals, average size, and range in size of grain.

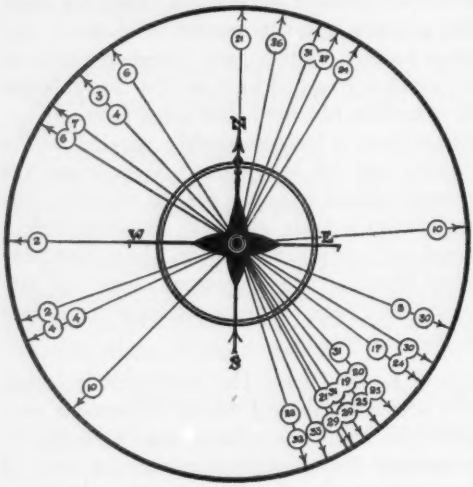

FIGURE 5.— *Relationship of ancient beach to floor of deposition.*

Discontinuous, nearly horizontal, sharply cross-laminated, sandy layers are interstratified with cross-laminated, micaceous layers of sandy silt. The cross-laminated sands are foreset laminae, dipping in many directions. The micaceous silts were evidently deposited in small, shallow basins, which were then partly eroded, and again, together with newly eroded basins, filled with sand and silt (Pl. 4, fig. 2). In every respect, therefore, these deposits seem to be strictly analogous to the modern beaches.

The preservation of a beach is manifestly contingent on its burial before erosion—a condition usually obtained by gradual submergence of the shore. As submergence of a beach progresses, the deposit of the lower foreshore overlaps that of the upper foreshore. Accordingly, superposition of a deposit of the lower foreshore on a deposit of the upper foreshore is to be expected in fossil beaches.

TERRACE DEPOSITS AT SAN PEDRO

At San Pedro, California, loosely consolidated, fossiliferous sands and gravels of the Pleistocene San Pedro formation rest on eroded edges of Miocene shale. The cross-laminated member of this formation is par-

ticularly well suited to this study, because the cross-lamination is excellently exposed in vertical section made by artificial excavation. This cross-laminated member lies below even-bedded layers of fossiliferous sand and rests upon strata of fossiliferous silt and dirty sand.

Uniformly sorted, medium-grained granitic sand, 80 per cent of which is between 0.25 and 0.5 millimeter in diameter, comprises the member. Thin, dark laminae, now considerably weathered (Pl. 7, fig. 1), contain grains of magnetite and dark-green hornblende. A few well-rounded cobbles, 2 inches in diameter, were found in this outcrop.

A surface of erosion, transecting the cross-laminated, sandy member, truncates laminae below and serves as a floor of deposition for laminae that gradually become tangent to this surface from above. In this exposure, the trace of the truncating surface normal to the strike of the laminae extends laterally for 185 feet. Although the average declivity of the total outcrop of the truncating surface is 6 degrees toward the sea, the slope is broken by a number of step-like terraces (Pl. 7, fig. 1).

The laminae below the truncating surface maintain an average dip of 16 to 18 degrees toward the present shoreline across the entire 185 feet of outcrop.

Sands above the truncating surface reveal two sets of cross-lamination: one, confined to the upper part of the outcrop; the other, to the lower part. The upper set of cross-lamination consists of discontinuous layers of laminated sand, 6 to 10 inches thick, each cross-laminated within its limiting surfaces by inclined laminae diverging in direction of dip. Limiting surfaces of single layers are discontinuous and approximately horizontal. The lower set of cross-lamination—that immediately above the truncating surface—has uncurved laminae, 8 to 10 feet long. The deposition, at slightly variant inclinations, of adjacent, uncurved laminae causes apparent curvature and reversal of dip of part of the lower set of laminae.

This sandy, cross-laminated member of the San Pedro formation, though it lies between fossiliferous, marine strata, contains no diagnostic fossils. Hence, reconstruction of the site of deposition of this cross-laminated member must depend almost entirely on the evidence of original structures.

The assumption that a beach was the site of deposition of this cross-laminated member of the San Pedro formation most satisfactorily accounts for the step-like terraces of the truncating surface, the peculiar reversal of dip in laminae immediately above the truncating surface, the discontinuous, cross-laminated layers separated by nearly horizontal surfaces, and the uniform sorting of materials. The step-like terraces of the truncating surface are comparable to those commonly formed on the

surface of a modern upper foreshore beach by changes in carrying power of waves at various heights of tide. The reversal of dip in long, nearly horizontal laminae characteristically takes place near the crest of a modern beach. The discontinuous, cross-laminated layers, separated by nearly horizontal surfaces, are similar to those that seemingly are forming on the lower foreshore at Newport Beach. These discontinuous, cross-laminated layers overlie the nearly horizontal laminae with reversed dips. This relationship of lower foreshore deposits superposed on upper foreshore deposits strengthens the supposition that the cross-laminated member is an ancient beach.

FIELD RELATIONS OF ANCIENT BEACHES

Thickness.—Ancient beaches characteristically are thin. This view is supported by both direct and indirect evidence. All the ancient beaches of this study are less than 40 feet thick, and the sandstone members of the Medina formation, which are considered beaches by several authors, are less than 25 feet thick.

Bailey Willis[6] infers that beaches are no thicker than the height of the waves that form them. This inference disregards eustatic changes in sea level, which, however, are carefully considered by Joseph Barrell.[7] Barrell deduces that: (1) beaches on emerged shores are washed away before being consolidated; (2) beaches on stationary shores of some relief may be buried by sediments from the land; (3) beaches on slowly submerging shores of low relief may escape destruction, if the beach is below the surface of planation of the transgressing sea. Hence, the thickness of ancient beaches is limited above by the depth of marine planation or by the height of waves, and below by the depth of the floor on which the beach is built. On a slowly subsiding shore of gentle relief, this depth is not great in most places.

Beaches that are low enough to escape destruction by marine transgression are subsequently buried by offshore marine sediment.

Continuity.—Ancient beaches in most places are not continuous at one horizon for great distances along the outcrop. The ancient beaches of this study thin rapidly and interstratify with marine or continental sediments.

Scarcity in Geologic Record.—Ancient beaches probably comprise only a small part of the total sedimentary record. This inference is supported by the following facts: (1) the volume of present beaches is small as compared with the total volume of all sedimentary deposits; (2) only a

[6] Bailey Willis: *Conditions of sedimentary deposition*, Jour. Geol., vol. 1 (1893) p. 436.

[7] Joseph Barrell: *Relative geological importance of continental, littoral, and marine sedimentation*, Jour. Geol., vol. 14 (1906) p. 444.

FIGURE 1. TRUNCATING SURFACE WITH BURIED TERRACE
Streaks are weathered, dark laminae. Terrace deposits at San Pedro, California.

FIGURE 2. TRANSVERSE VERTICAL SECTION OF BAYWARD SIDE OF SPIT
Bay to left. Entrance to Newport Bay, California.

DETAIL OF DEPOSITS

FIGURE 1. TRUNCATED EDGES OF LAMINAE IN LOBATE DUNE
Rogers Playa, Mohave Desert, California.

FIGURE 2. LOOSE SAND ON DUNE LAMINAE OF VARIANT DIP
Rogers Playa, Mohave Desert, California.

small part of the total volume of beaches is actually incorporated in the geologic record; (3) few ancient beaches were found by the writer in his search in many localities of California, though the sediments of the coast of California, which seem to have been subjected to several submergences and emergences, should contain more ancient beaches than most other localities, where recent eustatic changes in land and sea are less frequent.

Stratigraphic Relations.—Of the ancient beaches in California observed by the writer, one rests on a surface of erosion of considerable relief, which probably was cut into an old land surface by waves of a transgressing sea, and one rests on wind-laid sand.

Each of these ancient beaches is composed of two parts: a lower one, which has the characteristics of upper foreshore beaches; and an upper one, which has the characteristics of lower foreshore beaches.

At Newport Beach, the ancient lower foreshore beach is overlain by bay sediment; whereas, at San Pedro, the ancient lower foreshore beach is cut at the top by a surface of erosion.

At San Pedro, an advancing sea is indicated, inasmuch as wind-laid sand is buried by an upper foreshore beach, which, in turn, is buried by a lower foreshore beach. Probably, the superjacent surface of erosion, which truncates the lower foreshore beach, is an old surface of marine planation, in which erosion also removed any sediment that overlay the ancient lower foreshore beach.

The ancient beach at Newport Beach seems also to premise an advancing sea, which deposits the upper foreshore beach on the wave-eroded surface of an older landmass and the lower foreshore beach, in turn, on the upper foreshore beach.

ORIGINAL STRUCTURES OF OFFSHORE BARS, BARS, AND SPITS

ORIGINAL STRUCTURES OF BARS IN LAKE BASINS

The pioneer investigations of Gilbert [8] and Russell [9] indicate that bars (embankments) [10] in lake basins are composed of successive, partly eroded layers, which were deposited in the form of arches; these layers were limited by truncating surfaces of erosion, gently inclined near the crest but steepening landward. The landward limbs of these arches are steeper than the lakeward ones. Whether the crest of a bar shifts lakeward or landward depends on the level of water in the lake and on the nature of waves that swash the bar. When the level of water is higher than the bar, the crest is forced landward, and laminae are added to the landward

[8] G. K. Gilbert: *Lake Bonneville*, U. S. Geol. Surv., Mon. 1 (1890).

[9] I. C. Russell: *Geological history of Lake Lahontan, a Quaternary lake of northwestern Nevada*, U. S. Geol. Surv., Mon. 11 (1885).

[10] Embankment is Gilbert's term for undifferentiated bars, spits, and related shore forms.

side. When the level of water is lower than the bar, the active crest is moved lakeward, and laminae are added to the lakeward side. The shifting of the level of water effects a complicated internal structure, consisting of oppositely inclined laminae, which are limited by nearly horizontal surfaces of erosion and are interbedded with layers that show complete anticlines of deposition.

The writer has studied a few of these lake bars, and, in general, his observations tend to confirm those of Gilbert and Russell.

Inasmuch as changes in height of water over bars in lake basins resemble, except in duration, tidal changes in height of water over marine bars, original structures of lake bars are probably similar to those of marine bars.

ORIGINAL STRUCTURES OF MARINE OFFSHORE BARS, BARS, AND SPITS

Offshore Bars.—The term, offshore bar, designates the sandy deposit of large extent that forms some distance off shore on the gently sloping floor of the sea. Cross-sections of stone reefs of the Brazilian coast, which J. C. Branner [11] reports are consolidated offshore bars, have gentle seaward dips and structures of modern foreshore beaches. Inasmuch as the seaward side of an offshore bar is the functional equivalent of a beach, its structures correspond to those of a beach.

W. H. Twenhofel [12] points out that offshore bars, in their migration landward as a consequence of erosion on their seaward sides, are almost destroyed by activity of waves, but that laminae laid on the landward side of the crest of the bar are more likely to escape erosion.

Bars and Spits.—A great many bars, inasmuch as they begin as spits, should disclose, beneath the structure of the bar, the structure of a spit.

It is probably impossible to distinguish between littoral sandy deposits formed as bars and spits, especially after each has become a part of the stratified marine record, because the structures of each, and the processes forming these structures, are almost alike. The crest of a spit or bar, like that of a beach, migrates landward or seaward. Beyond the crest, on the side away from the sea, bars and spits are built into water, whereas beaches are built upon land. Spits, and most bars, too, in their early stages, grow at the distal end with addition of sediment supplied by littoral currents.

The writer's investigations of bars and spits in California indicate that the structure of the seaward side of a marine bar, or of a spit, is the same as that of a foreshore beach. The structure of the bayward side

[11] J. C. Branner: *The stone reefs of Brasil, their geological and geographical relations, with a chapter on the coral reefs*, Harvard Coll., Mus. Comp. Zool., Bull., vol. 44, Geol. ser. 7 (1904) p. 12.
[12] W. H. Twenhofel: *Treatise on sedimentation* (1932) p. 122. 2nd edition.

of an offshore bar, bar, or spit, which resembles that of a backshore beach, consists of layers of short, steeply inclined foreset laminae dipping bayward, interstratified with layers of long, gently inclined topset laminae, which also dip bayward (Pl. 7, fig. 2).

Inasmuch as the crest of a marine bar or spit migrates either landward or seaward, the crests of superjacent, normally partly truncated anticlines of deposition, which characterize the structure of a spit or bar, are staggered unevenly in vertical section. The seaward limb, composed of gently dipping, seawardly inclined, cross-laminated laminae, is opposed by the somewhat steeper, bayward limb of baywardly inclined, topset laminae, which limit layers of interstratified, short, steeply inclined, foreset laminae. Most laminae strike nearly parallel to the shoreline, but some of those of spits strike obliquely to the shoreline and dip in the general direction toward which the spit is being built.

ORIGINAL STRUCTURES OF DUNES

PRECEDENT INVESTIGATIONS

A review of the literature reveals that a group of criteria for recognizing deposits of eolian sand has been proposed. Huntington's [13] observation, that laminae of eolian stratification are tangent to the floor of deposition below, is accepted by Barrell,[14] Grabau,[15] and Cressey.[16] A variable direction of inclination of laminae is considered characteristic of dunes by Briart,[17] Grabau, Twenhofel, and Cressey. Twenhofel and Cressey find no horizontal laminae in dunes. According to Huntington, Barrell, and Grabau, foreset laminae of large scale are indicative of dunes.

Also suggested as criteria to recognize sands deposited by the wind are the observations of Kindle,[18] that the amplitude of wind-made ripples is less than that of water-made ones and that coarser sand rests on crests of wind-made ripples; and the observation of Dake,[19] that wind-laid grains manifest frosted surfaces.

DUNES NEAR ROGERS PLAYA, MOHAVE DESERT, CALIFORNIA

General Statement.—Internal structures of unconsolidated dunes are rarely exposed in cross-section, because loose sand immediately slides

[13] Ellsworth Huntington: *Some characteristics of the glacial period in nonglaciated regions*, Geol. Soc. Am., Bull., vol. 18 (1907) p. 380.

[14] Joseph Barrell: *Criteria for the recognition of ancient delta deposits*, Geol. Soc. Am., Bull., vol. 23 (1912) p. 434.

[15] A. W. Grabau: *A textbook of geology* (1920) p. 455.

[16] G. B. Cressey: *The Indiana sand dunes and shore lines of the Lake Michigan Basin*, Geog. Soc. Chicago, Bull. 8 (1928) p. 37.

[17] M. Briart: *Stratification entrecroisée*, Bull. Soc. géol. de France, 3d ser., vol. 8 (1879) p. 536.

[18] E. M. Kindle: *Recent and fossil ripple mark*, Geol. Surv. Canada, Mus. Bull. 25 (1917) p. 11.

[19] C. L. Dake: *The problem of the St. Peter Sandstone*, Missouri School of Mines and Metallurgy, Tech. ser., vol. 6 (1921) p. 180.

into a freshly made cut. Erosion of a group of semi-consolidated dunes on the northeast side of Rogers Dry Lake, Mohave Desert, California, reveals the structures of wind-laid sands in an intermediate stage between dunes and eolian sandstone. Remnants of semi-consolidated dunes, gullied by run-off of desert rains, are separated by broad, shallow, chute-shaped valleys, also wind-carved, which are partly filled with coarse, wind-rippled sand and gravel.[20]

Most of the material of the dunes is derived from Rogers Playa, which is flooded during the rainy season. The shores of this playa are composed of sand and gravel, but the central part consists, chiefly, of clay that has arrived in suspension in flood water. During the drying of the lake, the clay is deposited, so that the surface of the lake, when partly dried, appears as if coated with brown enamel. With further drying, the clay cracks, curls up, and is blown by wind across the smooth surface of the playa. While in transit, the chips of clay are broken and rounded to pellets of sand size, which, together with sand from the shores of the playa, provide the material for the dunes. The dune sand derived from the shores of the playa, is, for the most part, coarse, angular, and crudely sorted. Most grains, however, manifest surfaces that have been partly frosted and pitted. The pellets, when in place in the dunes, are disintegrated by percolation of rain water so that the clay is distributed and, when dry, makes an effective bond for the sand.

Stratification.—The entire dune is laminated. Laminae average 0.1 inch in thickness, but range from 0.05 inch to 4 inches. Many laminae are more than 50 feet long.

Coarser grades of sand are in gently inclined laminae that dip into the wind (windward laminae); finer grades, in the steeper, foreset laminae that dip with the wind (leeward laminae). In a few places, windward laminae of low dip rise to an untruncated summit curve and descend from the same curve to leeward laminae of steep dip. This structure is rare on a large scale, but is fairly common on a small scale. Where connected by a summit curve, the gradation in grain size, from windward to leeward laminae, is so gradual that it is almost imperceptible.

Laminae dip in almost every direction and with great variation in angle of dip. A statistical study of dips in typical layers of laminae reveals that steeply inclined leeward laminae have a prevailing component of easterly dip, whereas the gently inclined windward laminae have a prevailing component of westerly dip (Fig. 6). The steeply inclined laminae of the lee side of the dune dip away from the prevailing direction

[20] Excellent views of this area were published by Eliot Blackwelder: *Yardangs*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 159.

of wind; the gently inclined ones of the windward side, toward it. In general, steeply inclined laminae underlie those with gentle declivity.

Laminae on the lee side of a lobate dune are deposited in concentric shells, which are analogous to the beds of a plunging anticline, or, in the case of a barchan, to a plunging syncline. The truncated edges of such laminae, where eroded to a nearly horizontal surface, are concentric arcs (Pl. 8, fig. 1).

Laminae on the windward side of a dune probably have a similar distribution when an abundant supply of sand is available, but, with reduced supply or increased velocity, the wind scours broad, chute-like troughs between dunes and removes a large part of the windward laminae. Hence, outcrops of windward laminae are largely limited to elongate, sheet-like masses capping the crests of yardangs and to remnants that have escaped erosion along the sides and floors of chute-like valleys.

The wind-cut surface shown in Plate 8, figure 2, is partly covered with loose sand, which, if not removed, will be incorporated in the initial laminae of a new overlying layer—an illustration of one of the steps in the formation of an eolian sandstone.

Horizontal laminae are conspicuously absent.

Wind-cut surfaces that separate layers of wind-laid laminae are seldom continuous for great distances, because they are transected by other similar surfaces that have a different direction and angle of dip. Repetition of this transection of one surface by another results in truncating surfaces with limited area and with a variety of direction and angle of dip as great as that of the laminae. Such surfaces are almost never horizontal.

Interpretation of Stratification.—The characteristic extreme diversity in direction and angle of dip of laminae results from their deposition on the varied slopes of dunes; the diversity in direction and angle of dip of truncating surfaces results from erosion by wind of varied strength and direction on the slopes of these dunes; the arc-like arrangement of steeply dipping leeward laminae results from truncation of the leeward ends of laminated dunes, which may be either lobate or concave; the steeply inclined laminae, dipping approximately at right angles to the major wind direction, result from deposition on the sides of the wind-cut, chute-like valleys excavated from older dunes; the position of gently inclined windward laminae above truncated, steeply dipping, leeward laminae results from partial erosion of the windward side of the dune and its subsequent burial by windward laminae of the dune at a later stage, when its crest has migrated farther leeward; the relative coarseness of sand in windward, and fineness of sand in leeward, laminae is due to the carrying of the finer particles to the leeward side and the

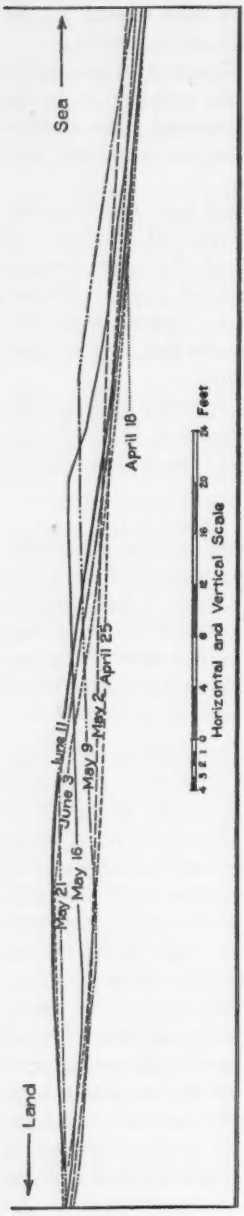

FIGURE 6.—*Distribution of dip of laminae in dunes*

Dips of groups of laminae in indurated dunes. Arrow indicates direction; encircled figure, the angle of dip. Rogers Playa, Mohave Desert, California.

leaving of the coarser, less easily transportable ones on the windward side of the dune.

CRITERIA FOR RECOGNITION OF EOLIAN SANDSTONE

The following original structures are probably characteristic of eolian sandstone: marked diversity in direction and angle of dip of laminae; relatively fine-grained, long, steeply inclined laminae arranged in concentric, either convex or concave, shells and overlain by elongate patches of relatively coarse, gently inclined laminae of opposite dip; few or no horizontal laminae; a marked divergence in direction and angle of dip of truncating surfaces in a small area; summit curves of small scale; and asymmetrical ripple marks of great wave length and small amplitude, with coarser sand on the crests than in the corresponding troughs.

Size and sorting of particles are not conclusive criteria by which to recognize an eolian sandstone. The results of the present study show that some dune sand is coarse and only crudely sorted, contrary to the view of some authors, that sands of eolian origin are fine-grained and uniformly sorted. On the other hand, frosting and pitting of grain surfaces seem to be valid criteria, inasmuch as many sand grains in the dunes of Rogers Playa, even though handled very little by the wind, are distinctly frosted and pitted. Tangential cross-lamination, which often has been cited as distinctive of eolian sandstone, is manifest in many sections of the indurated dunes at Rogers Playa. Random transverse sections through the leeward side of a

lobate dune expose laminae that steepen downward toward the floor of deposition; the upper parts of the laminae become tangent to overlying windward laminae, deposited on a truncated surface. In a barchan, cross-sections show leeward laminae tending to become tangent downward and to steepen upward. In straight-fronted dunes that have been partially eroded and subsequently buried by windward laminae, cross-sections show structures closely resembling torrential cross-lamination, except that the foreset and topset laminae of torrential cross-lamination dip in the same general direction, whereas those in dunes do not. It will be seen that the variety in direction of sections of dunes is almost unlimited and, hence, that the relations of laminae to both overlying and underlying truncating surfaces are almost unlimited.

SUMMARY AND CONCLUSIONS

Many of the original structures of beaches described in the literature are misleading because the authors fail to indicate to what part of the beach the structures belong.

The present paper describes structural differences between different parts of the beach and describes and illustrates, in detail, actual sections of modern beaches. These structures are compared with those of ancient beaches and with structures of bars and dunes that are likely to be confused with those of ancient beaches.

Although no single characteristic of beaches is likely to be distinctive enough to permit identification of an ancient beach in sedimentary formations, the combination of all, or many, of the structural features of beaches described in this paper should provide conclusive evidence of origin.

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GEOPHYSICAL INVESTIGATIONS IN THE EMERGED AND SUBMERGED ATLANTIC COASTAL PLAIN

PART I: METHODS AND RESULTS

BY MAURICE EWING, A. P. CRARY, AND H. M. RUTHERFORD

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INTRODUCTION

This report describes the development and application of a seismic method of procedure for making studies of submarine geology. Any new tool for gaining information about the submerged portion of the earth's crust would be noteworthy because few such tools are available. The present method merits additional consideration because its findings are directly translatable into such concrete terms as depth to certain horizons in the rocks and the elastic constants of the rocks.

In the seismic method, elastic waves, which are produced by means of explosives, travel through the upper layers of the earth to seismographs or geophones, which serve to measure the travel-time of the waves. In the seismic reflection method, the elastic waves studied are those that have gone almost directly downward and have returned to the surface as echoes from subterranean discontinuities. In the seismic refraction method, the waves studied are those that have penetrated to a zone in which their velocity is relatively high and have returned to the surface after traveling horizontally for considerable distance in this zone.

In the present extension of these methods of investigation to submarine problems, both the charge of explosive and the geophones were placed on the ocean floor, but in other respects the method of operation was similar to that used on land. Both reflection and refraction methods were used successfully, but the more important results came from the refraction measurements.

In its present state of development, the method is not applicable in water deeper than about 100 fathoms, or, in other words, beyond the edge of the continental shelf. For working in deeper water, additional seismic equipment and a ship capable of handling long heavy cables are needed. Preliminary designs for the necessary apparatus have been made. The equipment needed would be expensive, the demands upon operating technique great, and the rate of accumulation of data slow, but the results should be of sufficient interest to justify the effort.

The locations at which measurements were made are shown in Figure 1. In addition to four stations at sea south of Woods Hole, Massachusetts, a fairly complete section across the submerged and emerged portions of the Atlantic Coastal Plain at Cape Henry, Virginia, have been studied by the refraction method. These studies classified the subsurface materials into three zones: unconsolidated sediments, semi-consolidated sediments, and crystalline rocks; both the thickness and the velocity of longitudinal elastic waves in each zone were determined.

Seismic reflection tests were made at scattered points along the Cape Henry section, and numerous reflections of good quality were obtained. As the reflection stations were spaced several miles apart, it has been impos-

sible to correlate the various reflecting horizons from station to station. However, the reflection studies have definitely proved that good reflections

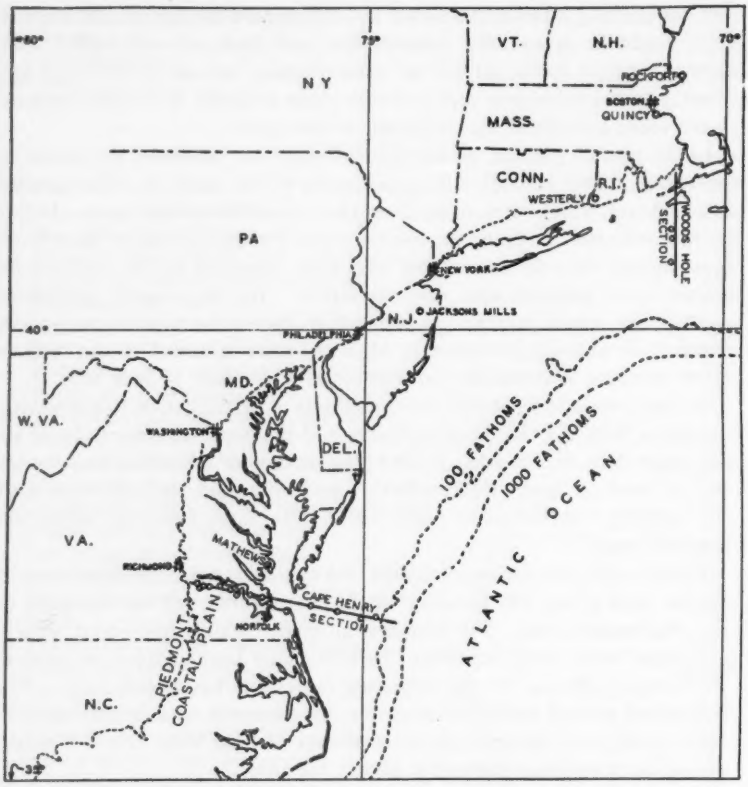

FIGURE 1.—*Key map of regions studied*

may be obtained in several hundred feet of water, and they have given qualitative corroboration of the depth to crystalline rocks—a determination deduced from refraction tests.

ACKNOWLEDGMENTS

The writers are indebted to the Geological Society of America, the Woods Hole Oceanographic Institution, and the United States Coast and Geodetic Survey for generous and substantial contributions of funds and facilities, without which the project could not have been undertaken.

The willingness of officials of Lehigh University and the University of Pittsburgh to grant leave of absence for making the cruise on the *Atlantis* is greatly appreciated.

The Geophysical Research Corporation loaned the geophones used in the work and gave valuable cooperation in several other instrumental problems.

C. O. Iselin and H. C. Stetson gave much assistance in perfecting the technique of refraction measurements, and the officers of the *Oceanographer* and of the *Atlantis* made every effort to facilitate the work. William Bowie, N. H. Heck, R. M. Field, and B. L. Miller gave valuable assistance and encouragement at every stage in the development of the project.

HISTORICAL BACKGROUND

The earliest published discussion of the feasibility of studying submarine geology by the seismic method is probably that by DeGolyer[1] in the symposium on the "Application of Geophysics to Ocean Basins and Margins," held at the annual meeting of the American Geophysical Union in 1932. Since then, the application of the seismic method has been prominent among the projects receiving attention of the Committee on Geophysical and Geological Study of Oceanic Basins,[2] of the American Geophysical Union, whose chairman, R. M. Field, has been an able leader in the study of submarine geology.

Reflection and refraction measurements were made in water-covered areas as early as 1927 for the purpose of locating oil-bearing structures.[3] Ewing and Rutherford took part in a survey of the coastal lakes of Louisiana in 1928. The experience gained in working in water 5 to 20 feet deep proved valuable in the present project. In the previous work, the geophones were planted securely in lake-bottom mud by the use of long poles (a procedure not feasible at sea); the recording instruments were carried in launches capable of executing maneuvers impossible for ships to perform; and distances were obtained by measuring travel-times of air waves. Hence, it seems reasonable to classify the measurements made in the lakes and bayous of the Gulf Coast in the same group with land operations, and to consider the present work the initial application of seismic measurements to submarine geological problems.

The field measurements herein described were made by the present writers, who had the benefit of the advice of B. L. Miller on geological

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[1] Everett DeGolyer: *The application of seismic methods to submarine geology*, Am. Geophys. Union, Tr. (1932) p. 37.

[2] R. M. Field: *Report of committee on geophysical and geological study of oceanic basins*, Am. Geophys. Union, Tr. (1933) p. 13; (1935) p. 6; (1936) p. 8.

[3] E. E. Rosaire and O. C. Lester, Jr.: *Seismological discovery and partial detail of Vermillion Bay salt dome, Louisiana*, Am. Assoc. Petrol. Geol., Bull., vol. 16 (1932) p. 1221.

questions. The first preliminary tests were made in May 1935, in the vicinity of the deep bore-hole at Jackson Mills, New Jersey (Fig. 1).

The next tests were made between June 17 and June 30, 1935, on board the ship *Oceanographer,* of the United States Coast and Geodetic Survey, which was engaged in hydrographic work off Cape Henry. It was a great privilege to be accompanied on this cruise by Captain N. H. Heck, Chief of the Division of Seismology and Terrestrial Magnetism of the United States Coast and Geodetic Survey, whose broad experience in scientific work at sea enabled him to give much valuable advice and assistance.

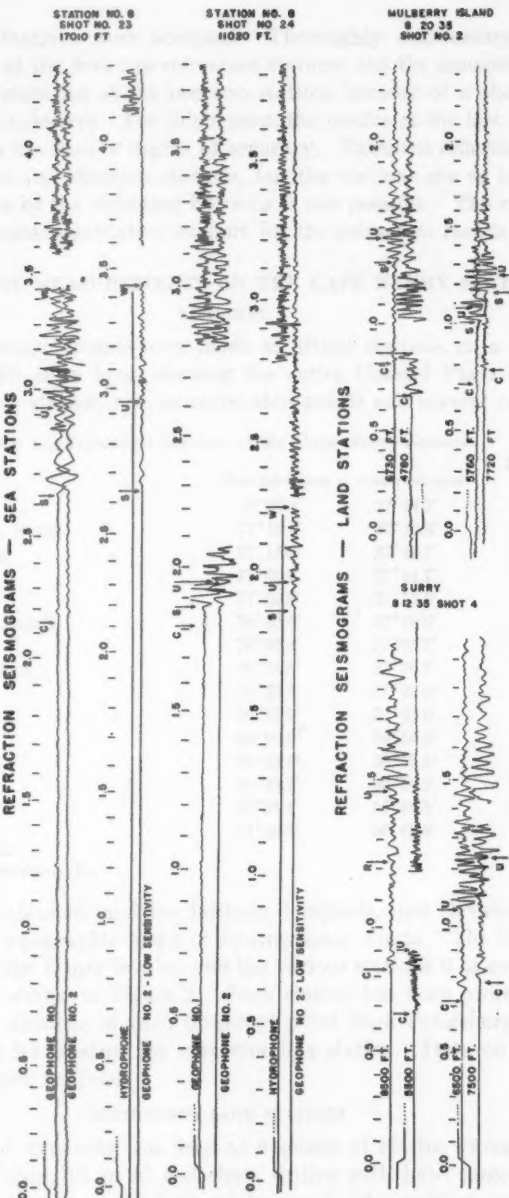
The work on the *Oceanographer* proved the feasibility of the project. It was confined to reflection tests, the most successful of which were made at night, when the ship was anchored.

Calibration tests on land by the reflection method were made from July 2 to 15 at Mathews and at Fort Monroe, Virginia, near deep wells, which extend to the crystalline rocks. It soon became apparent that one, or more, long refraction profiles were needed, and by July 25, arrangements were completed for seismic refraction work. The transmission of seismic energy by the sediments of the region was found to be phenomenally efficient, so that, although the supply of explosives was limited, many more refraction measurements could be made than had been anticipated. As refraction measurements give more complete information than do reflection measurements, and as the former are not dependent upon calibration of velocities at a point where the depth is known, the feasibility of using this type of measurement was of the greatest importance.

July 25 to August 22 was spent in making seismic refraction measurements at eleven stations between Cape Henry and Petersburg, Virginia. These formed a vital part of the work. They gave ample calibration as well as definite information about the conditions to be expected at the first stations at sea, and reduced the whole problem of working at sea to a gradual progression from the known to the unknown.

The final measurements at sea were made on the ship *Atlantis,* of the Woods Hole Oceanographic Institution. A short cruise on the *Atlantis* was made October 16 to 19, during which time, refraction measurements were made at four stations, on a line extending south from Woods Hole. These were the first refraction tests to be made at sea, and, during the development of the technique, the writers had the assistance of C. O. Iselin and H. C. Stetson, of the Woods Hole Oceanographic Institution, and of R. M. Field, of Princeton University. The data for these stations are incomplete, but, nevertheless, they afford important confirmation for certain results on the Cape Henry profile.

Actual measurements for the sea stations on the Cape Henry profile occupied the time from October 26 to 30. Four refraction stations and



REPRESENTATIVE SEISMOGRAMS

six reflection stations were occupied. Thoroughly satisfactory results were obtained at the first two refraction stations, but the amount of testing had to be curtailed at the last two stations because of a shortage in the supply of explosives. For this reason, the results at the last two stations are below the desired degree of accuracy. Excellent reflections were obtained at the six reflection stations, but the stations are so far apart that correlation of the reflecting horizons is not possible. The reflection data offer valuable qualitative support for the refraction results.

REFRACTION MEASUREMENTS ON THE CAPE HENRY SECTION

LOCATION

Refraction measurements were made at fifteen stations, on a line approximately 150 miles long, crossing the entire Coastal Plain at Cape Henry. At each station, one, or more, shot points and several recording

TABLE 1.—*Refraction Stations of the Cape Henry Section*

	Station	West Longitude	North Latitude	Distance * (miles)
1	Camp Lee	77°20.3′	37°14.7′	3.4
2	Youngblood's Store	77°19.6′	37°10.8′	4.6
3	Prince George	77°15.0′	37°11.7′	9.0
4	Newville	77°08.9′	37°11.7′	14.4
5	Brandon	77°02.6′	37°12.9′	20.1
6	Mercy Seat Church	76°52.6′	37°08.9′	29.8
7	Surry	76°46.4′	37°05.7′	35.9
8	Mulberry Island	76°34.8′	37°06.7′	46.7
9	Hampton	76°23.9′	37°03.6′	57.0
10	Fort Monroe	76°17.9′	37°01.0′	62.7
11	Cape Henry	76°03.9′	36°54.2′	77.9
12	Station 8	75°43.2′	36°57.3′	95.3
13	Station 12	75°25.7′	36°54.2′	112.0
14	Station 16	75°01.1′	36°49.5′	135.4
15	Station 20	74°49.6′	36°49.8′	145.9

* Miles east of Petersburg, Va.

points were established, and the latitude, longitude, and elevation were obtained from topographic maps or hydrographic charts. The line representing the Cape Henry Section and the various stations that constitute the section are shown in Figure 2. Each station has been projected to this line. The distance of each projected point from Petersburg serves as a coördinate for locating the corresponding station. Data on station locations are given in Table 1.

METHOD FOR LAND STATIONS

Fifty per cent dynamite was used as a source of elastic waves. This was placed in holes, 10 to 15 feet deep, drilled with hand augers. An insulated wire, forming part of the plate circuit of a radio transmitter,

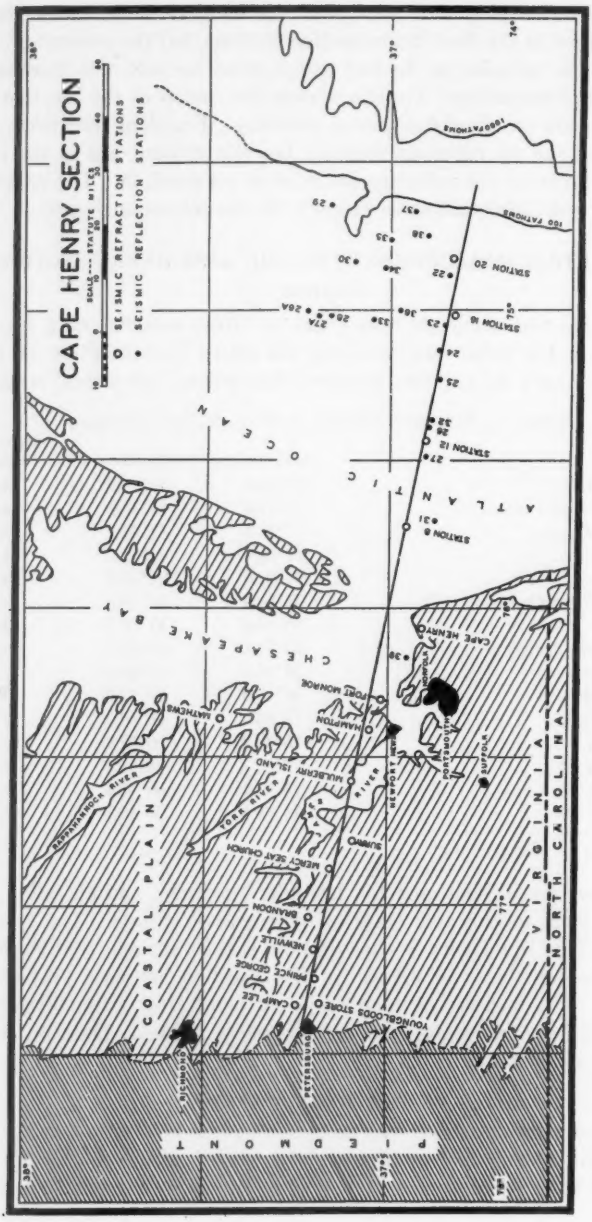


FIGURE 2.—*Map of the Cape Henry section*

passed down the hole and through the charge of dynamite. Shortly before the charge was to be fired, the key of the transmitter was depressed, initiating a signal, which was terminated when the insulated wire was broken by the explosion.

The seismograms were recorded photographically on two three-element, hand-cranked oscillographs (donated by the Geophysical Research Corporation), which were interconnected by a belt to produce an improvised six-element instrument. Each oscillograph contained a 50-cycle tuning fork, which placed time marks on the seismograms at intervals of one hundredth second and served as a standard for measurement of the travel-time. The oscillograph elements had natural frequencies of 90 cycles per second and were approximately critically damped in oil. The geophones used to detect earth vibrations consisted essentially of permanent magnets, mounted on spring suspensions, which permitted motion in the vertical direction only; they were arranged to induce currents in coils attached to the framework. The geophones were adjusted to have natural frequencies of 35 cycles per second and were critically damped in oil. A four-stage resistance-capacity coupled amplifier was used in connection with each of the geophones.

Four geophones, placed at recording points usually spaced at 1000-foot intervals, were used. The radio signal from the shot point, denoting the instant at which the explosion occurred, was recorded photographically in both oscillographs. Where the points were all on a single straight line, distances between shot and recording points were measured by chaining for stations; otherwise, the distances were determined by a transit and stadia rod survey.

At each station, travel-times were measured for several different distances approximately on a straight line, and time-distance graphs were drawn. From these graphs, calculations of velocity and depth were made. Details for individual stations are given in later sections.

METHOD FOR SEA STATIONS

At sea, a blasting gelatin, sometimes used as an oil-well explosive, designed to function effectively under great hydrostatic pressure, was used. The charge was prepared by lashing together the required number of sticks, priming one stick with a waterproof electric blasting cap, weighting the bundle with sash weights, and lowering it to the bottom by means of the firing cable. Ordinarily, these charges were not placed in the water more than 15 minutes before firing. All were prepared, loaded, and fired from a small whale boat, which was at times as far as 8 miles from the ship carrying the recording apparatus. The skill and seamanship of the officers and men of the *Atlantis*, together with remarkably calm weather, which prevailed during the tests, made it possible to obtain

results by this method without taking dangerous chances, but it would be decidedly hazardous to attempt it under less favorable circumstances.

An auxiliary blasting cap, connected in series with the one used as a detonator, served to break a wire in the plate circuit of the transmitter to produce the time-signal. (This cap was placed in air behind a suit-

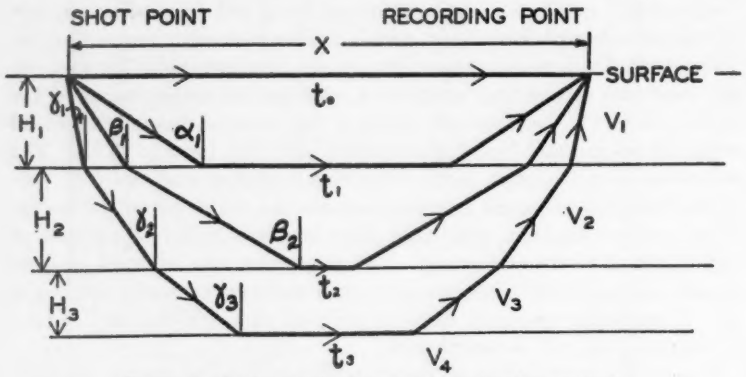


FIGURE 3.—*Paths of seismic waves for refraction method*

able shield, which prevented damage from flying bits of copper.) An excessive current was used to fire the two caps in order to insure against errors in timing, which would be introduced by failure of simultaneous detonation.[4] This arrangement for producing the time signal was selected to avoid the necessity of running wires of the timing circuit to the submerged charge.

Two geophones were used; after the ship was at anchor, they were lowered to the bottom by means of the cables that connected them with the recording apparatus. Suitable bases prevented their overturning. The two geophones were at essentially the same distance from the shot and gave essentially the same data. The use of duplicate geophones and recording systems was definitely worthwhile, because the two records obtained on each shot served to check each other and made simple the problem of distinguishing true ground motions from stray vibrations transmitted down the connecting cables. Where the ship was rolling badly, the geophones were lowered from opposite sides of the vessel, so that one geophone was certain to be undisturbed at any moment, even though the other might be rendered inoperative by a pull on the cable. As an anchored ship usually swings through a considerable arc, it was

[4] G. H. Loving and G. H. Smith: *Explosives and electric blasting caps for geophysical prospecting*, Soc. Petrol. Geophys., Jour., vol. 6 (1935) p. 27.

necessary to organize the work so that the shot could be fired soon after the geophones had reached bottom.

To facilitate the identification of certain waves, one geophone was arranged to give a low-magnification seismogram in addition to the usual high-magnification one. It permitted identification of the wave that travelled through the water. This wave was preceded by one or more waves through the ground. The water wave had a distinctive high-frequency character, which was not present elsewhere on the record. Although the water wave could be readily identified on the low-magnification seismogram, it was considered advisable to remove the possibility of uncertainty in identification by placing a hydrophone a few feet beneath the water and connecting it to record the arrival of the water wave on the seismogram.

The distance for each shot was derived from the travel-time of the water wave, and the general method of measuring the time may be considered comparable to the accurate system of radio acoustic ranging used at present by the United States Coast and Geodetic Survey.[5] Readings of temperature and salinity of the sea water at four depths at each station were taken by A. H. Woodcock, for the purpose of determining the velocity of sound in the water. The results give a mean value of 4949 ± 5 feet per second. A velocity of 4950 feet per second was adopted for calculating all distances. The water wave afforded an accurate and convenient means of measuring the distance.

PRINCIPLES OF INTERPRETATION

The standard method of interpreting the data of the refraction seismograph was used; for convenience, it will be outlined here. Consider the subsurface to be composed of a number of zones, separated from each other by horizontal planes. Let the velocities of seismic waves in the various zones be V_1, V_2, and so on; the thicknesses of the various zones, h_1, h_2, and so on; the travel-times for the various possible paths as shown in Figure 3, t_1, t_2, t_3, and so on. It may be seen that

$$t_1 = x/V_1$$
$$t_2 = 2h_1 \cos \alpha_1/V_1 + x/V_2$$
$$t_3 = 2h_1 \cos \beta_1/V_1 + 2h_2 \cos \beta_2/V_2 + x/V_3$$
$$t_4 = 2h_1 \cos \gamma_1/V_1 + 2h_2 \cos \gamma_2/V_2 + 2h_3 \cos \gamma_3/V_3 + x/V_4$$
$$\sin \alpha_1 = V_1/V_2$$
$$\sin \beta_1 = V_1/V_3, \sin \beta_2 = V_2/V_3$$
$$\sin \gamma_1 = V_1/V_4, \sin \gamma_2 = V_2/V_4, \sin \gamma_3 = V_3/V_4.$$

[5] N. H. Heck, E. A. Eckhardt, M. Keiser: *Radio acoustic method of position finding in hydrographic surveys*, U. S. Coast and Geodetic Surv., Spec. Publ. 107 (1924).

Putting $x = 0$ in these equations and putting T_1, T_2, T_3 for the corresponding values of time, one obtains:

$$T_1 = 2h_1 \cos \alpha_1 / V_1$$
$$T_2 = 2h_1 \cos \beta_1 / V_1 + 2h_2 \cos \beta_2 / V_2$$
$$T_3 = 2h_1 \cos \gamma_1 / V_1 + 2h_2 \cos \gamma_2 / V_2 + 2h_3 \cos \gamma_3 / V_3.$$

These formulas suffice for all calculations on the refraction results made in this report. The travel-time curve represented by these equations consists of several straight lines, or branches, as illustrated in Figure 28. Each path shown in Figure 3 corresponds to a branch of the travel-time curve, and each branch of the curve to a definite subsurface zone. The four zones with which the present paper is concerned will be called the weathered, the unconsolidated, the semi-consolidated, and the crystalline zones. These zones, and also the phases and the branches of travel-time curves associated with them, will be designated by the letters W, U, S, and C, respectively. A phase is a definite arrival of energy as recorded on a seismogram. A phase is said to be normal provided there is some distance at which it will be the first phase to arrive; otherwise, it is said to be masked. For example, the S phase is masked in Figure 24 and is normal in Figure 28. The reciprocal slope of a branch of the travel-time curve gives the velocity of seismic waves in the corresponding subsurface zone. Thus, the values of h_1, h_2, . . . are obtained by substitution in the formulas of the values of V_1, V_2, . . . and T_1, T_2, . . . which are read from the travel-time curves.

Where the boundary plane between two zones is inclined to the surface, a slight modification of the formulas is required; it is also necessary to gather additional data to detect and measure the slope. The method usually used for slopes, known as "reversed profiles", was used on the Surry station and is illustrated in Figure 16. Here, the travel-time to each recording point was measured from two different shot points, one at each end of the line. The travel-times from the two shot points are plotted separately, and a slope in the subsurface layers would be revealed by failure of corresponding branches of the two travel-time curves to show equal velocities. The slope in the surface of the crystallines at the Surry station is seen to be negligible.

A procedure designed to reveal the presence of a slope, without entailing additional observations, was used on most of the land stations. It is well illustrated by the Camp Lee profile (Figs. 4 and 5). The four geophones were placed at stations 1000, 1500, 2000, and 2500 feet along the line of profile. Shots were fired at stations 3000 and 5000, and the eight travel-times obtained from these two shots are plotted on the single curve of Figure 4. The absence of an appreciable slope is shown by the fact that the C branch of the travel-time curve is a single line.

A slope in the surface of the crystalline rocks would cause the four points derived from the shot at 5000 to lie on a line parallel to, but either above or below, the curve determined by travel-times from the shot at 3000. It should be noted that this method of procedure actually simplified field operations, for the shot point could be moved more

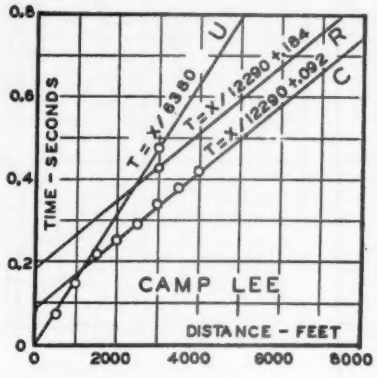


FIGURE 4.—*Travel-time curve for Camp Lee station*

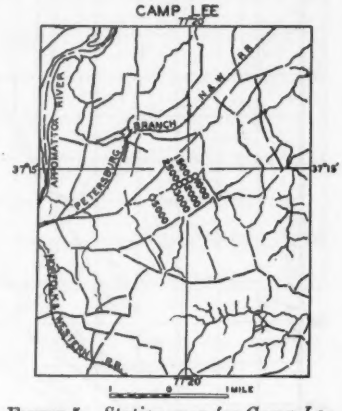


FIGURE 5.—*Station map for Camp Lee*

conveniently than could the geophones, cables, and recording apparatus. At no station was a slope detected.

LATER PHASES

In general, impulses travel from each shot to a recording point along several of the paths sketched in Figure 3. As these impulses usually arrive at different times, each seismogram records several different impulses or phases. "Later phases" are impulses that arrive at the recording point subsequent to the arrival of one, or more, waves. A later phase is recognized on a seismogram by a change in amplitude, frequency, or phase of the recorded motion. Frequently, later phases show only vaguely on seismograms, so that they cannot be identified with the desired degree of certainty. In the present work, certain later phases were so phenomenally distinct that they could be identified positively and timed precisely. Most of the travel-time curves bear many points, which are derived from later phases, especially on the U branches. This favorable circumstance permitted the necessary data to be obtained from a much smaller number of shots than would have been otherwise needed.

On practically every profile the U phase appeared with great amplitude, even at distances so great that two other phases were arriving

ahead of it. From Mulberry Island to Cape Henry the S phase was masked, and appeared only as a later phase. When the data for these four profiles were first obtained, the existence of the S zone had not been

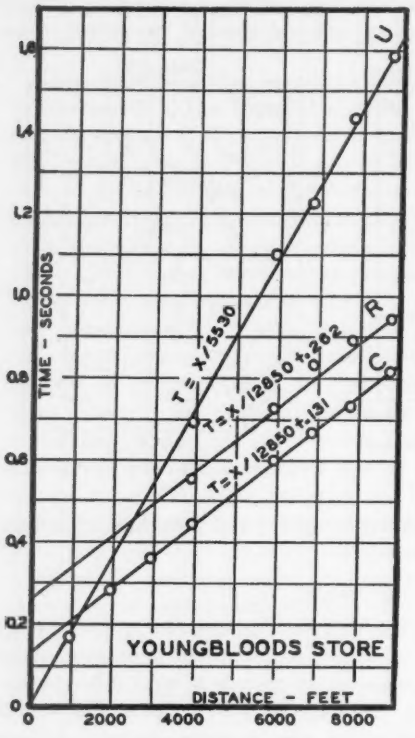


FIGURE 6.—*Travel-time curve for Youngbloods Store station*

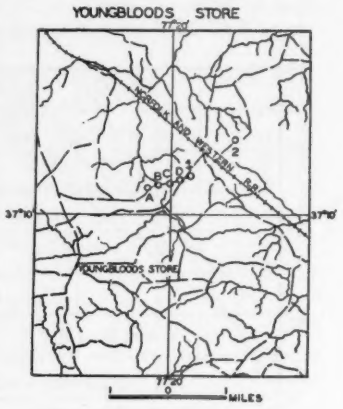


FIGURE 7.—*Station map for Youngbloods Store*

anticipated. The points for the travel-time curves were plotted as they now appear in Figures 18-24, and those points that are now attributed to the S phase were then very puzzling. Several months later, the data for the adjacent sea stations were obtained. The travel-time curves for stations 8 and 12 reveal the semi-consolidated zone at a depth so slight, compared to that of the crystallines, that the S phase is no longer masked and is the first to arrive over considerable distances. This establishes, beyond question, the existence of the semi-consolidated zone and indicates clearly the proper interpretation for the adjacent land stations.

STATIONS

Camp Lee.—Figures 4 and 5 show the travel-time curve and the sta-

tion map for Camp Lee. This is a good measurement, in which the first arrivals are all sharp and clear. For intermediate distances, the U

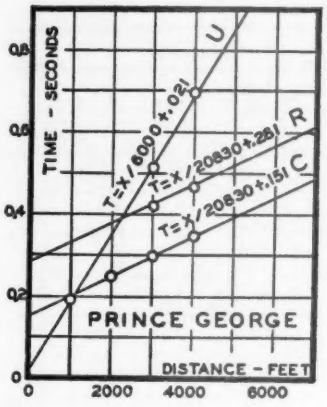


FIGURE 8.—*Travel-time curve for Prince George station*

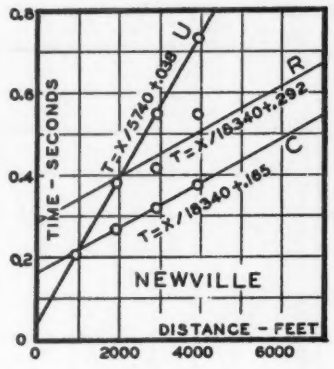


FIGURE 10.—*Travel-time curve for Newville station*

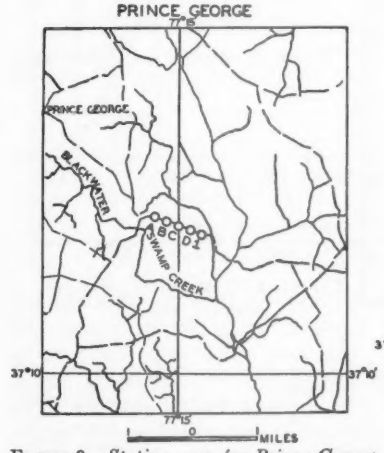


FIGURE 9.—*Station map for Prince George*

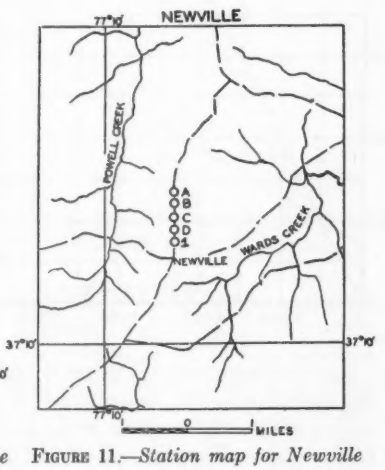


FIGURE 11.—*Station map for Newville*

phase could not be read, because the amplitude of motion on the seismogram was too great. From Figure 31, it is seen that the velocity 12290 ft/sec, obtained here for the crystalline rocks, is far below the normal value.

As is true for all travel-time curves in this report, the lines drawn on the graph and the equations given for these lines were determined by the method of least squares. The travel-time data, calculations, and results for all profiles are given in Tables 2 and 3.

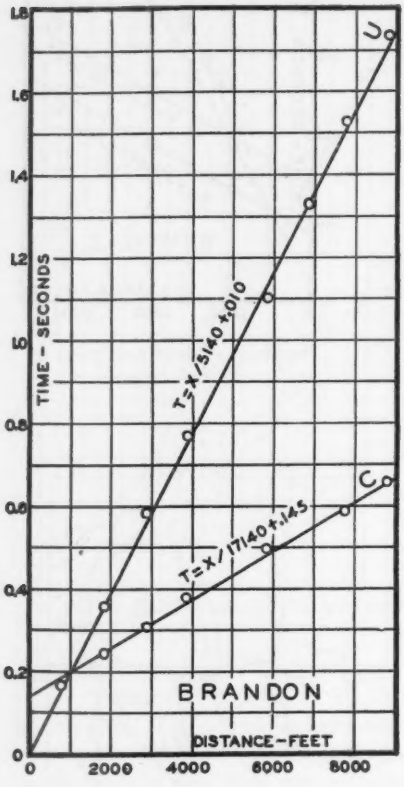


FIGURE 12.—*Travel-time curve for Brandon station*

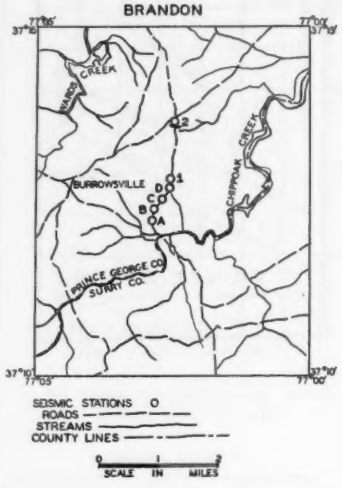


FIGURE 13.—*Station map for Brandon*

Youngbloods Store.—Figures 6 and 7 show the travel-time curve and the station map for Youngbloods Store. This is a good measurement, having sharp first arrivals and excellent definition of the U branch by later phases. The low velocity of 12850 ft/sec for the crystallines appears here, as at Camp Lee. It is certain that the bedrock here is different than at other places studied in this section. Miller suggests that a buried Triassic Basin may be present beneath these two stations.

A later phase, appearing on the records of this station, is designated as R on the travel-time curve. The line for this phase has the same

slope and double the intercept of the C line. It corresponds to waves that have been reflected once from the surface during their progress from the shot to the recording station. Frequently, it has been considered that

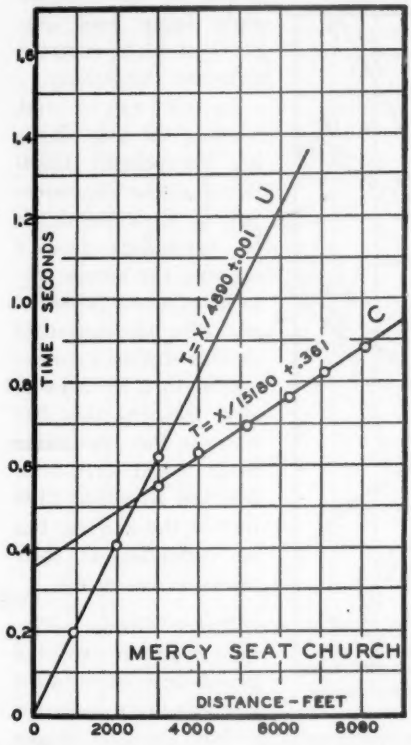


FIGURE 14.—*Travel-time curve for Mercy Seat Church station*

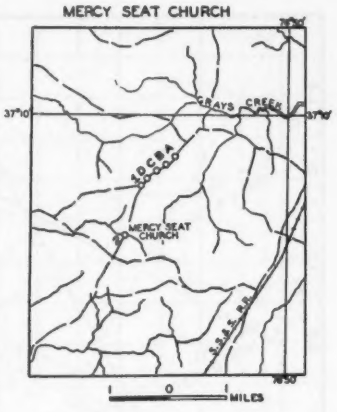


FIGURE 15.—*Station map for Mercy Seat Church*

multiple surface reflections prove that the paths along which the energy travels are curved by refraction, as described by Ewing, in an earlier report,[6] but such is not the case. An interpretation of this reflected phase, on the basis of straight line paths like those of Figure 3, is made, when it is recognized that, although the straight-line system causes the energy reflected at a given surface point to spread out continuously before returning again to the surface, it compensates for this effect by providing a continuous set of points at which the surface reflection may occur. In other words, it is not necessary that the point of reflection of a surface-reflected wave lie midway between the point of

[6] Maurice Ewing and L. D. Leet: *Seismic propagation paths*, Am. Inst. Min. Eng., Tr., Geophys. Prosp. (1932).

origin and the point of reception. Thus, the surface-reflected phase, sometimes considered to be the crucial test, leads to no conclusion whatever in the question of curved paths versus straight ones. The seismograms of this station show fairly good evidence of waves reflected twice at the surface.

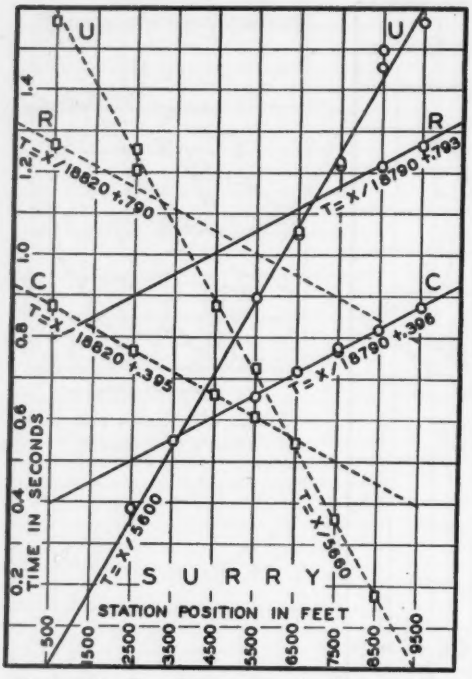


FIGURE 16.—*Travel-time curve for Surry station*

In order to test whether the zone showing the velocity 12580 ft/sec might be underlain at moderate depth by crystalline rocks showing the normal velocity of about 17800 ft/sec, the measurements on this station were extended to a distance of 9000 feet. At this distance the crystalline rocks would have been detected if within 2100 feet of the surface, but no indication of their presence was found.

Prince George.—Figures 8 and 9 show the travel-time curve and the station map for Prince George. This is a good measurement, in which the later phases determine the velocity in the unconsolidated, well enough to permit a depth measurement with only one shot. As only one shot was made, it is impossible to tell whether the observed velocity of 20830 ft/sec represents the true velocity in the crystallines or whether it is altered by a slope in the surface of the crystallines. As no slope was found at any of the stations where the measurements were so arranged that they would have detected it, the most reasonable conclusion is that the observed value of velocity indicates a change in the composition of the rock.

The seismograms of this station clearly exhibit the R phase.

Newville.—Figures 10 and 11 show the travel-time curve and the station map for Newville. In this measurement, the U and C phases are well developed to give a good determination of depth, but no estimation

of slope is possible. The velocity, 18300 ft/sec, obtained here is considered to be the normal value for the crystalline rocks of the entire Cape Henry section.

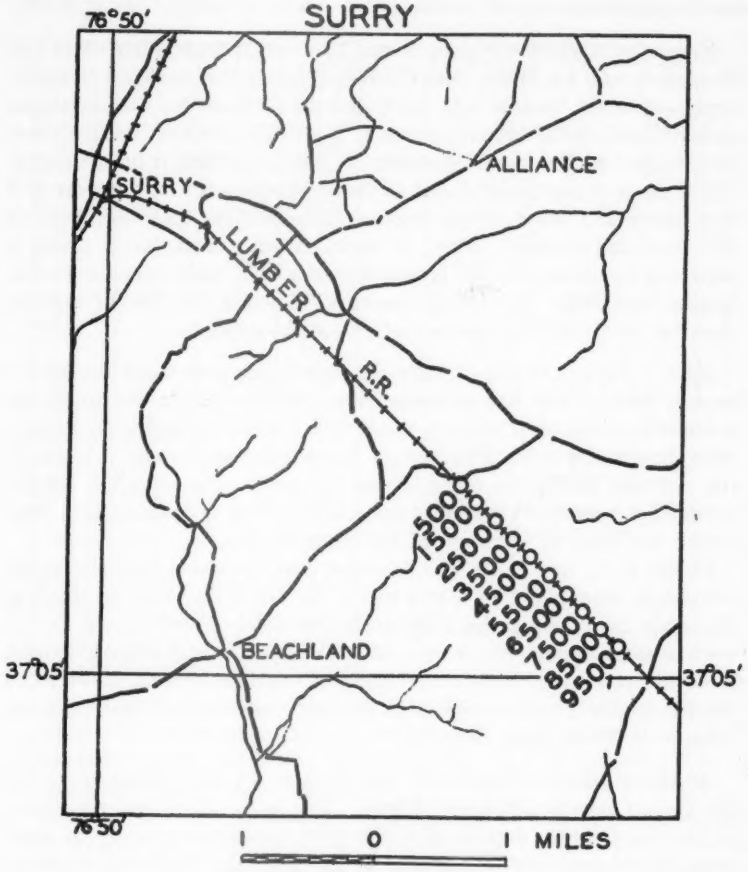


FIGURE 17.—*Station map for Surry*

The R phase looks good on the seismogram, but, as seen from the travel-time curve, the time of arrival of the phase is not in good agreement with the expected value.

Brandon.—Figures 12 and 13 show the travel-time curve and the station map for Brandon. This measurement is satisfactory. The U phase is particularly well developed, and the C phase is acceptable, despite the

fact that on the longer distances the amplitude of earth motion was low enough to impair the precision. No slope is indicated, and the velocity 17100 ft/sec, found for the crystallines, is normal. The R phase could not be identified.

Mercy Seat Church.—Figures 14 and 15 show the travel-time curve and the station map for Mercy Seat Church. The control on depth measurement is excellent, because both the U and the C phases are well developed as first phases. The reflected phase R is probably present, but it cannot be identified with sufficient certainty to justify plotting it on the graph. The U phase is poorly developed at distances where it would appear as a later phase and shows characteristics different from those exhibited at any preceding station. There is slight evidence that the S phase is beginning to appear, but the erratic behavior of U makes definite identification impossible. The travel-time curve indicates that the low velocity observed for C probably was caused by a slight slope.

Surry.—Figures 16 and 17 show the travel-time curves and the station map of Surry. The line of measurement of this station lies along the grade of an abandoned railroad, which runs straight through an unusually level stretch of woods. This station was selected as the one at which to run reversed profiles, in order to test for slope. The velocities for the crystalline from the two profiles agree well enough to indicate zero slope within the limit of accuracy of the measurement.

Phases U, C, and R are well recorded, and the travel-time curves are accurately determined. On two traces the amplifiers were arranged so that only high frequencies (100 cycles, or higher) were recorded, but, even in these cases, both earlier and later phases were readily identified. No evidence of semi-consolidated material could be seen on the excellent records obtained, so it is concluded that semi-consolidated material is not present at Mercy Seat Church.

Mulberry Island.—Figures 18 and 19 show the travel-time curve and the station map for Mulberry Island. This is the most westerly station of the Cape Henry Section that definitely shows the presence of semi-consolidated material. For several months after the data for this station, and for the land stations east of it, were obtained, it seemed impossible to get an interpretation that did not contradict at least one rather well-established fact. When the results from the sea stations became available, the existence of the semi-consolidated zone became obvious. Inclusion of the zone in the calculations for the last four land stations cleared up immediately all difficulties of interpretation.

It should be noted that in the last four land stations the S branch of the travel-time curve is masked—that is, it never appears as a first

arrival. This circumstance is due to the presence of the crystalline zone, in which the velocity is high, at a depth not sufficiently in excess of that of the semi-consolidated material to prevent masking by the waves through the crystalline zone. The masking disappears for the sea sta-

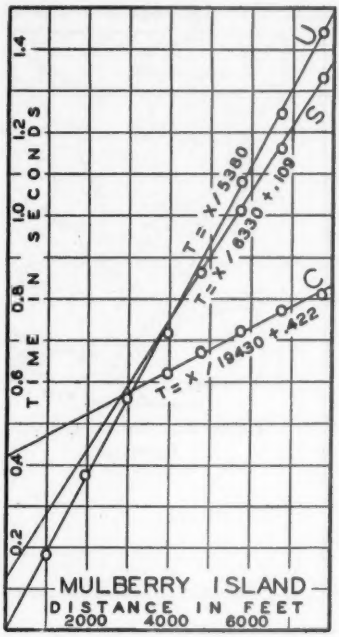


FIGURE 18.—*Travel-time curve for Mulberry Island station*

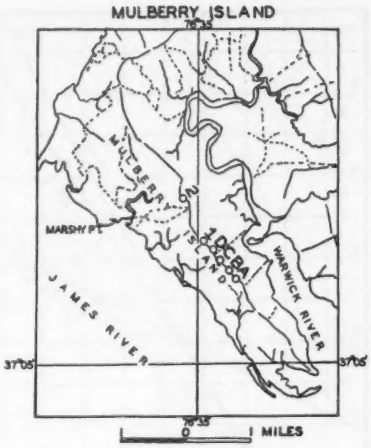


FIGURE 19.—*Station map for Mulberry Island*

tions, simply because the thickness of the semi-consolidated material is greater at those stations. Allen [7] has discussed a similar situation. With reference to some travel-time curves, which he has worked out for a hypothetical structure, he states:

"It is apparent that the wave from the third horizon would never be the first impulse, since the intersection of the T_2 and T_3 curves is past the point where the T_4 wave arrives first. The only manner by which the third horizon velocity may be determined is to select the secondary waves on records shorter than 4.174 kilometers, which is the value of x where the T_2 and T_4 curves intersect in this particular case."

Allen fails to mention the possibility that his "masked" T_3 waves may be picked as second arrivals beyond the point where they cross the T_2 line—the most likely of the two possibilities, except in cases further complicated by abnormal absorption of elastic energy. At any rate, the

[7] T. L. Allen: *Use of record character in interpreting results and its effect on depth calculation in refraction work*, Am. Assoc. Petrol. Geol., Bull., vol. 16 (1932) p. 1212.

phases used in the present work to map the masked horizon are quite definite and give a satisfactory solution to the problem.

The velocity found here for the crystalline is considered to be normal and not appreciably influenced by slope.

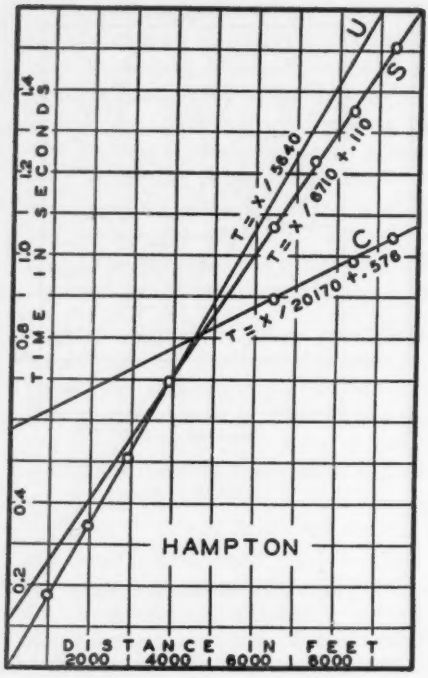


FIGURE 20.—*Travel-time curve for Hampton station*

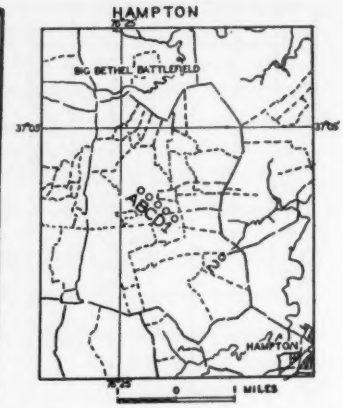


FIGURE 21.—*Station map for Hampton*

Hampton.—Figures 20 and 21 show the travel-time curve and the station map for Hampton. The measurements for this station followed a line of the Chesapeake and Potomac Telephone Company, through a dense wood, and the writers are indebted to E. G. Magruder, of that company, for assistance in locating the station on the map. The seismic data are satisfactory but are not exceptionally precise. All three branches of the travel-time curve are well determined, but it was impossible to pick out the waves through the unconsolidated zone as third arrivals at the greater distances. The presence of the semi-consolidated zone is clearly shown.

Fort Monroe.—Figures 22 and 23 show the travel-time curve and the station map for Fort Monroe. This is a poor measurement. Ground

unrest, caused by running machinery and traffic, difficulty in sinking a hole of suitable depth for burying the charge of explosive, and other hindrances imposed by the geography of the location made it impossible to get seismograms which could be read with high accuracy. Further, only

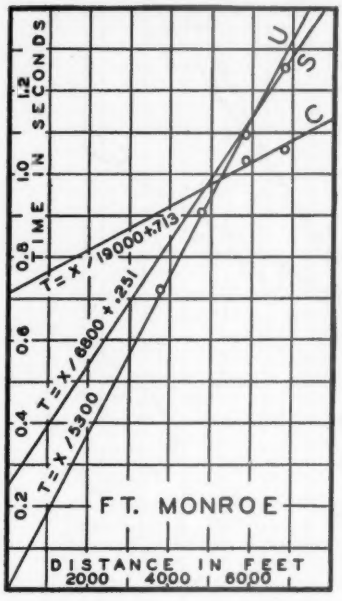


Figure 22.—*Travel-time curve for Fort Monroe station*

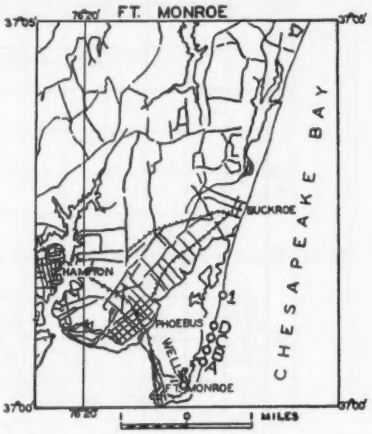


Figure 23.—*Station map for Fort Monroe*

one shot was made, and the data are, therefore, meager and do not suffice to give velocities in the semi-consolidated and crystalline zones at this point. For calculating the depths, the velocities for the semi-consolidated and the crystalline zones were taken as 6800 ft/sec and 19000 ft/sec, respectively, by interpolation between the values found for adjacent stations. From the composite velocity diagram of Figure 31, it may be seen that the interpolation is valid. Although the data available at this station would not be adequate for a satisfactory depth determination if the station stood alone, a reliable calculation becomes possible when the interpolations are made. The depth to the crystalline rocks, calculated from the seismic measurements, is 2260 feet. The depth to crystalline rocks found in the deep well drilled inside Fort Monroe in 1902 is 2243 feet.[8] The agreement is perhaps partly fortuitous, and the error less than the probable error of the seismic determination, in view of the poor quality

[8] Samuel Sanford: *Underground water resources of the coastal plain province of Virginia*, Va. Geol. Surv., vol. 5 (1913) p. 99.

of the data available. There would be a large discrepancy between seismic result and well log, if the S phase, which is masked here, had been

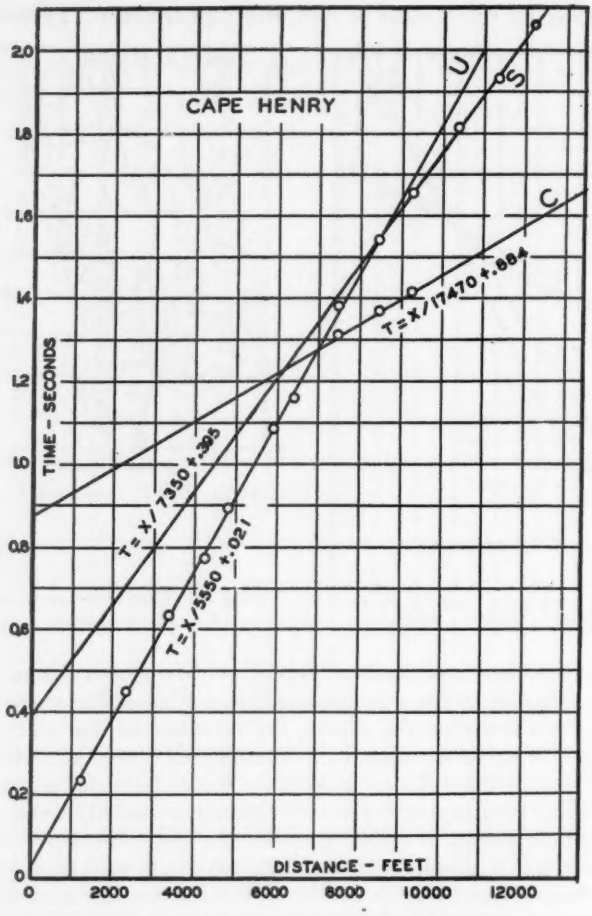


FIGURE 24.—*Travel-time curve for Cape Henry station*

overlooked. The well is located almost a mile south of the center of the line on which the seismic measurements were made (Fig. 23).

Cape Henry.—Figures 24 and 25 show the travel-time curve and the station map for Cape Henry. The branches of the travel-time curve for

the unconsolidated and semi-consolidated zones are well determined. The branch for the crystalline is determined with only moderate accuracy,

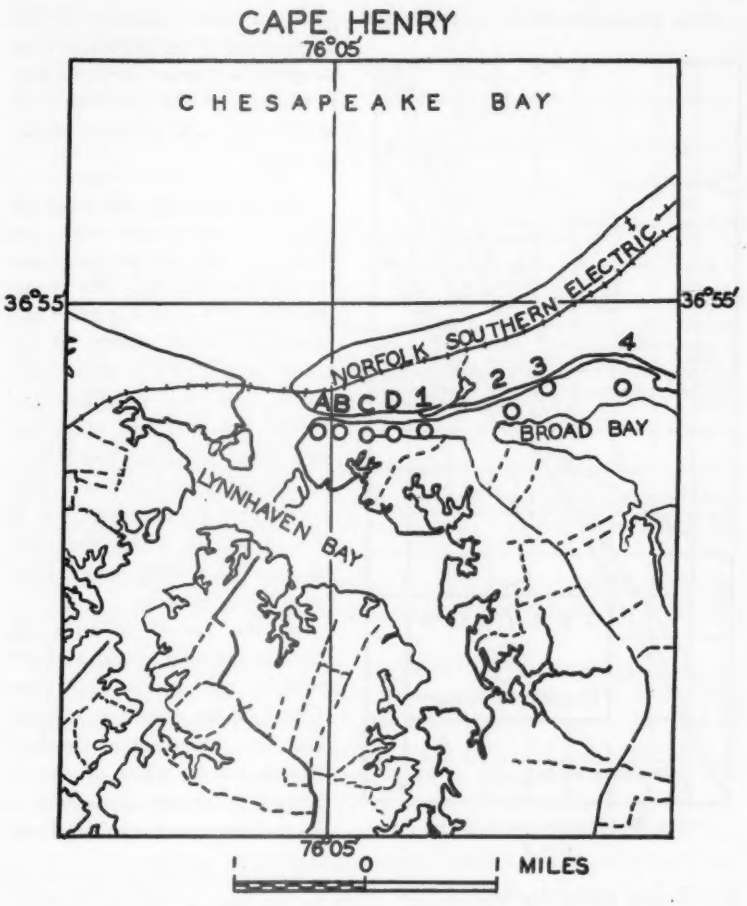


FIGURE 25.—*Station map for Cape Henry*

because the first arrivals on the seismograms at the greater distances were low in amplitude. It was not feasible to get more amplitude, because the explosive charges could not be buried deeply enough for effective performance in the sand dune terrain of this locality. With the hand augers used for sinking the shot holes, it was impossible to go beneath the water table.

Even though the shot points were on the highest sand dunes available, the 12 or 15 feet of loose sand lying above the shot did not provide sufficient confinement.

This measurement is reliable and gives an ideal example of the application of the refraction seismograph to places where one zone is masked by the presence of an underlying zone of much higher velocity.

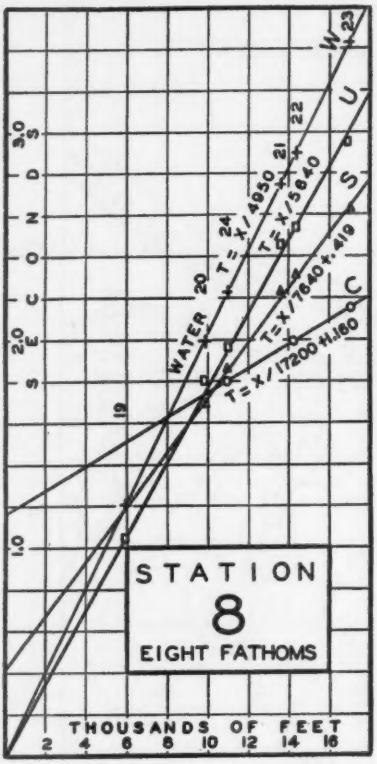


FIGURE 26.—*Travel-time curve for Station 8*

No. 8.—Figures 26 and 27 show the travel-time curve for Station 8 and the station maps for all sea stations. The measurement at Station 8 is quite satisfactory. Each branch of the travel-time curve is well determined, and the velocities and depths obtained are consistent with those found at Cape Henry, the adjacent land station. This is the first station described in this report for which the wave through the semi-consolidated zone is not masked.

On the record of Shot 20, there is a feeble phase at 1.609 seconds, which has not been plotted on the graph. It is likely that this phase is a "leader", similar to those which will be described under the discussion of the measurements at Matthews.

No. 12.—Figure 28 shows the travel-time curve for Station 12. This is the best of the sea stations. The U phase is picked on the low-magnification record. It is an angular impulse, larger in amplitude than any arriving before it, which may be correlated from shot to shot with certainty. Study of the high-magnification records shows that this phase might be picked a trifle earlier, but the low-magnification readings are accepted because of the definite nature of the correlation which may be made from them. No essential variation in interpretation is involved.

The S phase is characterized by a gradual building up to a regular sinusoidal wave of rather low frequency. It is readily identifiable by its character. This phase is unaccountably almost 0.2 second early on Shot 31.

The C phase is of low amplitude on the three longest shots, but it is definite. The first phase of Shot 29 is intentionally used for both the C and the S branches of the curve, on the basis of record character.

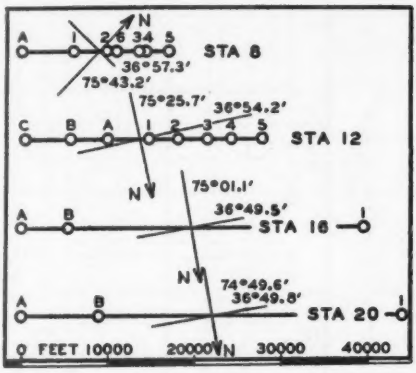


FIGURE 27.—*Station map for Stations 8, 12, 16, and 20*

No. 16.—Figure 29 shows the travel-time curve for Station 16. This determination is incomplete, because both the number of shots and the amount of explosive used on each shot had to be curtailed because of a shortage of the dynamite supply. It is decidedly unfortunate that more data could not have been obtained here, because a few shorter shots would be of great value in determining the S phase more accurately.

The U phase may be considered to be completely determined by the data at hand. The value obtained is consistent with those for preceding stations.

The velocity for S was taken, by extrapolation from Figure 31, as 8180 ft/sec, and a line of this slope was drawn between the two points available for fixing the position of this branch of the curve. Because of the poor control of the S branch, the depth given for the semi-consolidated zone must be considered no better than a rough approximation.

No direct evidence, whatever, was obtained for the depth to the crystalline zone, so nothing more than a minimum limit for the depth to it can be set. The velocity of 18000 ft/sec for C was taken by extrapolation from Figure 31. The record for Shot 32 is poor, so that one may not be certain whether or not a C phase arrived ahead of the S. A good record was obtained from Shot 33, and it is certain that no phase arrived ahead of the one plotted. Therefore, the depth given for the crystalline zone is not an observed depth but, rather, the minimum depth that is consistent with the observed data.

No. 20.—Figure 30 shows the travel-time curve for Station 20. The data here are more complete than at Station 16 but are still decidedly

inadequate, because of the shortage of the explosive supply. The U phase is satisfactory. The velocity 8390 ft/sec for S, obtained by linear extra-

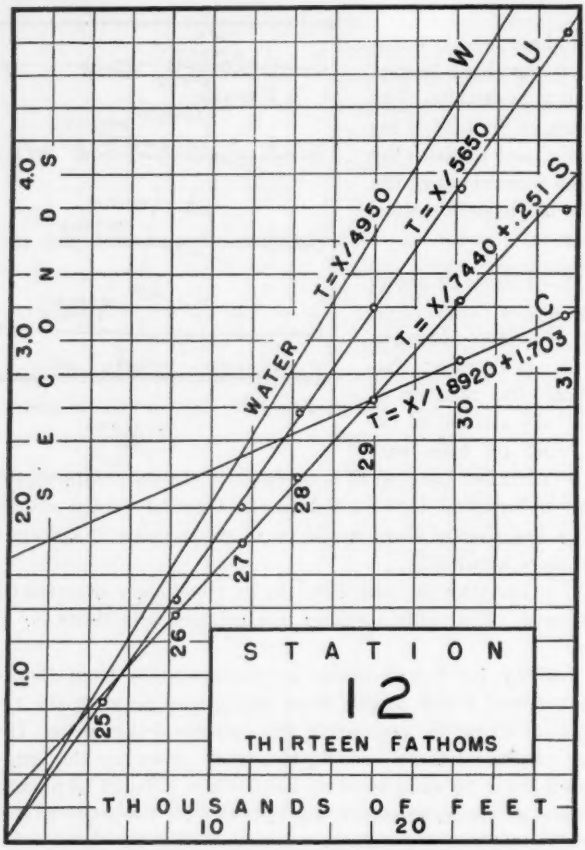


FIGURE 28.—*Travel-time curve for Station 12*

polation from Figure 31, is in good agreement with the two points available on the travel-time curve, but these points afford poor control on the calculation of the depth to the semi-consolidated zone. The velocity 18000 ft/sec for the C phase is, likewise, obtained by extrapolation, but the point on the travel-time curve through which the C line is drawn is quite definite on the seismogram. The depth to the crystalline zone obtained here is, therefore, reliable insofar as the basic assumption—that the three

zones revealed by the stations to the west persist here—is justified, for which there is considerable evidence from the seismograms.

For Shot 35, the geophones were on the bottom, in 46 fathoms; they functioned perfectly at that depth. The depth of 35 fathoms given on Figure 30 refers to the average depth over the 8-mile line covered by the measurements for this station.

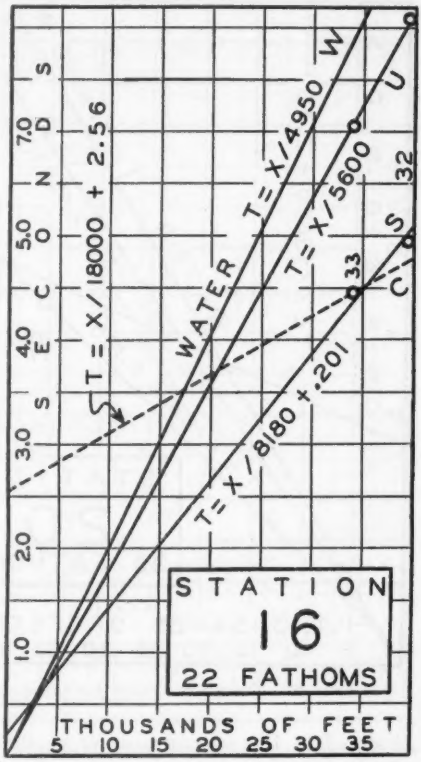


FIGURE 29.—*Travel-time curve for Station 16*

VELOCITY ANALYSIS

Figure 31 shows the velocities found for the three zones, for the entire Cape Henry section. All values obtained by interpolation are enclosed in dotted circles.

The velocity for U remains reasonably constant and has no uniform trend. The curve for S shows a uniform trend, the velocity increasing to the east. The straight line shown through these points was used as the basis for the interpolation. Toward the west, the velocity for S becomes so low that it is difficult to distinguish this phase from U. This accounts for S being terminated rather abruptly at the Mulberry Island Station (Fig. 32).

The velocities of C fluctuate considerably, probably because of the effects of slopes, which were not completely eliminated by the method of controlling slope used in the work. The only variation of velocity to which significance is attached is that found at Camp Lee and at Youngbloods Store, which has been mentioned as indicating a Triassic basin.

RECORD CHARACTER

As a rule, those phases on seismograms that represent refracted waves have gradual beginnings. Even though large charges of explosives are

used, it is often difficult to cause refracted waves to begin sharply enough for satisfactory timing. Contrary to this rule, all waves refracted through the crystalline rocks at land stations in the present work show sharp beginnings, having a characteristic wave-shape resembling the letter W. Following the large amplitude beginnings, there is little motion on the record until the S and the U phases arrive. The C phases were so small at the sea stations that it cannot be stated with certainty that they possess the characteristic shape.

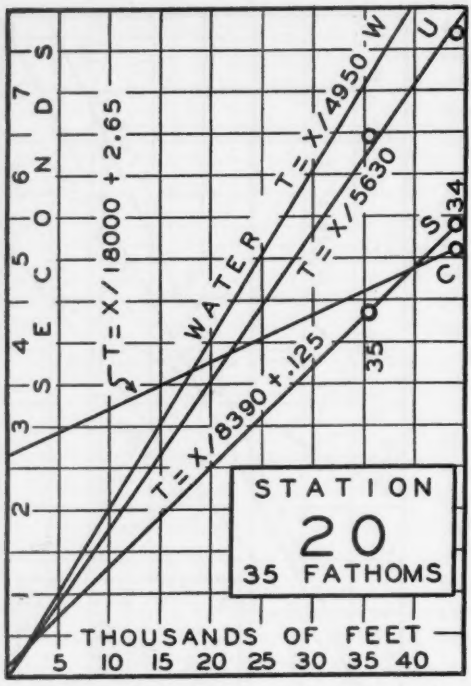


FIGURE 30.—*Travel-time curve for Station 20*

Records from land stations show prominent "ground roll". This consists of a long train of waves—period 0.06 to 0.13 second, velocity 800 to 2000 ft/sec, amplitude large—that are usually considered to be similar to the Rayleigh waves and the Love waves observed in the seismograms of earthquakes. These waves are completely absent at sea stations, and this feature is the outstanding point of difference between the two sets of records. Even in the earlier portions of the seismograms, the frequencies of the waves at sea stations are higher—roughly twice as high—than those for land stations. Plate 1 shows tracings of some refraction seismograms, which illustrate some of these features.

SUMMARY

Table 2 shows a summary of all the travel-time data for the stations of this section. For the sea stations of this section, the distances were obtained by multiplying the travel-time of the water wave, indicated by *w* in Table 2, by 4950 ft/sec. The shot locations and recording locations are on the station maps.

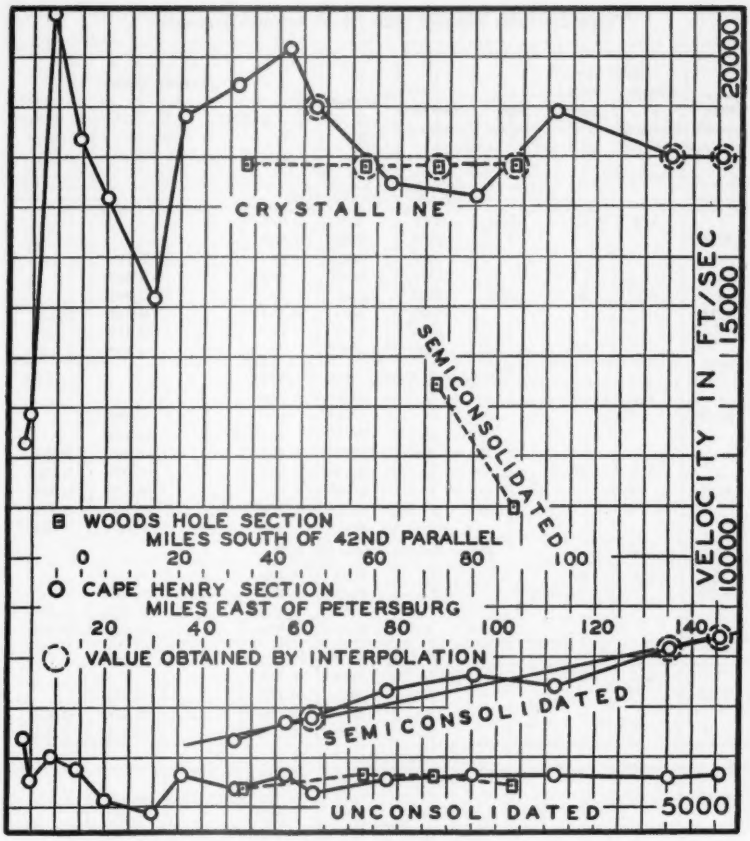


FIGURE 31.—*Velocity chart for Cape Henry and Woods Hole sections*

Table 3 summarizes the calculations and results. All values of velocity and time shown in this table were taken from the various travel-time curves. The values of h refer to the thicknesses of the various zones and were calculated by means of the formulas given on p. 763. The elevations given in the last three columns are taken with reference to sea level and refer to the upper surfaces of the zones for which they are given.

Figure 32 shows the section across the coastal plain, which has been constructed from the seismic measurements already described. The uniform elevation of the surface of the crystalline zone at the five most westerly stations should be noted. Possibly, the change from this uniform elevation is more abrupt than is shown in the section. It should be re-

TABLE 2.—*Time distance data for all refraction stations*

Shot number	Charge of explosive (lb.)	Shot hole depth (ft.)	Shot location	Recording location	Distance X (ft.)	Travel Time U (sec.)	Travel Time S (sec.)	Travel Time C (sec.)	Travel Time R (sec.)
Camp Lee, August 14, 1935									
1	3.5	15	3000	2500	500	0.071	.....	.....	.....
8/14				2000	1000	0.148	.....	.....	.....
				1500	1500	os	.....	0.218	.....
				1000	2000	os	.....	0.252	.....
2	6.0	20	5000	2500	2500	os	.....	0.290	.....
8/14				2000	3000	0.474	.....	0.338	0.429
				1500	3500	os	.....	0.378	os*
				1000	4000	os	.....	0.417	os
Youngbloods store, August 17, 1935									
1	3.5	10	1	D	1000	0.169	.....	.....	.....
8/17				C	1985	os	.....	0.283	.....
				B	2976	os	.....	0.360	.....
				A	3970	0.694	.....	0.444	0.558
2	10.0	17	2	D	5956	1.099	.....	0.600	0.730
8/17				C	6896	1.225	.....	0.670	0.833
				B	7833	1.433	.....	0.733	0.892
				A	8784	1.585	.....	0.815	0.947
Prince George, August 16, 1935									
2	3.5	10	1	D	1000	0.192	.....	.....	.....
8/16				C	2000	0.352	.....	0.247	.....
				B	3000	0.513	.....	0.295	0.419
				A	4000	0.694	.....	0.343	0.466
Newville, August 16, 1935									
1	3.5	12	1	D	1000	0.214	.....	.....	.....
8/16				C	2000	0.388	.....	0.275	.....
				B	3000	0.554	.....	0.328	0.425
				A	4000	0.730	.....	0.384	0.559
Brandon, August 15, 1935									
2	.....	.....	1	D	767	0.167	.....	.....	
8/15				C	1813	0.355	.....	0.249	
				B	2868	0.582	.....	0.305	
				A	3871	0.769	.....	0.379	
1	6.0	17	2	D	5821	1.100	.....	0.496	Low
8/15				C	6797	1.336	.....	.....	ampli-
				B	7757	1.525	.....	0.587	tude
				A	8786	1.731	.....	0.658	
Mercy Seat Church, August 18, 1935									
1	.....	.....	1	D	965	0.202	.....	.....	.....
8/18				C	2052	0.413	.....	.....	.....
				B	3045	0.627	.....	0.556	.....
				A	4055	poor	.....	0.636	.....
2	.....	.....	2	D	5232	.....	.....	0.702	.....
8/18				C	6247	poorly	.....	0.774	poorly
				B	7133	devel-	.....	0.831	devel-
				A	8101	oped	.....	0.892	oped

* Off-scale on the seismogram.

TABLE 2.—*Time distance data for all refraction stations—continued*

Shot number	Charge of explosive (lb.)	Shot hole depth (ft.)	Shot location	Recording location	Distance X (ft.)	Travel Time			
						U (sec.)	S (sec.)	C (sec.)	R (sec.)
Surry, August 12–13, 1935									
1	5.5	10	500	1500	1000	.....	.....	.....	.....
8/12				2500	2000	0.384	.....	.....	.....
				3500	3000	0.551	.....	.....	.....
				4500	4000	.....	.....	.....	.....
4	.....	.....	500	5500	5000	0.896	.....	0.659	.....
8/12				6500	6000	1.059	.....	0.719	.....
				7500	7000	1.226	.....	0.767	.....
				8500	8000	1.454	.....	0.816	1.218
1	.....	.....	500	6500	6000	1.052	.....	0.716	.....
8/13				7500	7000	1.220	.....	0.773	.....
				8500	8000	1.498	.....	0.828	.....
				9500	9000	1.564	.....	0.874	1.268
3	.....	.....	9500	8500	1000	0.177	.....	.....	.....
8/12				7500	2000	0.361	.....	.....	.....
				6500	3000	0.543	.....	.....	.....
				5500	4000	0.725	.....	0.608	.....
2	.....	.....	9500	4500	5000	0.877	.....	0.661	.....
8/12				3500	6000	.....	.....	.....	.....
				2500	7000	1.257	.....	0.766	1.203
				1500	8000	.....	.....	.....	.....
				(500)	9000	1.564	.....	0.874	1.268
Mulberry Island, August 19–20, 1935									
1	.....	.....	1	D	1000	0.183	.....	.....	.....
8/20				C	1985	0.387	.....	.....	.....
				B	2980	0.561	.....	.....	.....
				A	3979	0.718	.....	0.621	.....
2	.....	.....	2	D	4780	.....	0.866	0.672	.....
8/20				C	5758	1.082	1.012	0.723	.....
				B	6731	1.244	1.165	0.774	.....
				A	7725	1.440	1.331	0.812	.....
Hampton, August 3–5, 1935									
1	4.0	15	1	D	975	0.177	.....	.....	.....
8/3				C	1950	0.343	.....	.....	.....
				B	2925	0.510	.....	.....	.....
				A	3900	0.697	.....	.....	.....
1	9.0	.....	2	D	6442	.....	1.065	0.897	.....
8/5				C	7417	.....	1.228	.....	.....
				B	8392	.....	1.350	0.988	.....
				A	9367	.....	1.507	1.043	.....
Fort Monroe, August 21, 1935									
1	.....	10	1	D	3766	0.721	.....	.....	.....
8/21		bay		C	4789	0.909	.....	.....	.....
				B	5848	1.095	.....	1.033	.....
				A	6811	.....	1.253	1.060	.....

TABLE 2.—*Time distance data for all refraction stations—continued*

Shot number	Charge of explosive (lb.)	Shot hole depth (ft.)	Shot location	Recording location	Distance X (ft.)	Travel Time			
						U (sec.)	S (sec.)	C (sec.)	W (sec.)
Cape Henry, August 6–7, 1935									
1			1	D	1222	0.235	.....	.....	.....
8/6				C	2332	0.449	.....	.....	.....
				B	3357	0.634	.....	.....	.....
				A	4253	0.772	.....	.....	.....
2			2	D	4803	0.893	.....	.....	.....
8/6				C	5899	1.085	.....	.....	.....
1	16.5		3	D	6397	1.159	.....	.....	.....
8/7				C	7476	1.379	.....	1.311	.....
				B	8446	1.539	.....	1.370	.....
2			4	D	9256	.....	1.654	1.413	.....
8/7				C	10360	.....	1.811	.....	.....
				B	11350	.....	1.931	.....	.....
				A	12220	.....	2.062	.....	.....
Station 8, October 26, 1935									
19	1.5	.....	1	A	5960	1.046	.....	.....	1.205
20	2	.....	2	A	9860	1.799	1.696	.....	1.993
21	5	.....	3	A	13600	2.454	2.224	.....	2.748
22	5	.....	4	A	14340	2.539	2.309	1.993	2.808
23	15	.....	5	A	17010	2.950	2.624	2.148	3.436
24	10	.....	6	A	11020	1.956	1.862	1.800	2.226
Station 12, October 27, 1935									
25	2	.....	1	A	4780	0.840	.....	.....	0.966
26	3	.....	2	A	8250	1.455	1.361	.....	1.666
27	5	.....	3	A	11530	2.008	1.793	.....	2.329
28	6	.....	4	A	14350	2.570	2.199	.....	2.809
29	8	.....	5	A	17920	3.215	2.650	2.650	3.620
30	15	.....	5	B	22240	3.906	3.245	2.879	4.494
31	20	.....	5	C	27440	4.852	.....	3.153	5.543
Station 16, October 28, 1935									
32	20	.....	1	A	39520	7.09	4.95	.....	7.983
33	30	.....	1	B	34080	6.06	4.45	.....	6.884
Station 20, October 28, 1935									
34	25	.....	1	A	44030	7.68	5.39	5.10	8.894
35	25	.....	1	B	35500	6.46	4.34	.....	7.172
Woods Hole, October 16, 19, 1935									
2		.....	4	B	3520	.....	.....	.286	.711
3		.....	5	B	1980	.....	.....	.185	.400
16		.....	1	A	1170	.231	.....	.154	.....
		.....	1	B	570	.110	.....	.....	.....
17		.....	2	A	2560	.....	.....	.228	.518
		.....	2	B	1940	.....	.....	.194	.391
18	3	.....	3	A	4140	.....	.....	.307	.836
		.....	3	B	3590	.665	.....	.291	.725
Station 1, October 17, 1935									
5		.....	3	A	2260	.408	.....	.....	.456
		.....	3	A′	2360	.433	.....	.....	.476
6		.....	2	A	3300	.581	.....	.....	.667
7	2	.....	1	A	4360	.756	.....	.655	.881
Station 3, October 18, 1935									
13	5	.....	1	A	11090	1.932	1.343	.....	2.241
14	6	.....	2	A	9460	1.763	.....	.....	1.911
15	6	.....	2′	A	9260	1.589	1.207	.....	1.871

TABLE 2.—*Time distance data for all refraction stations—continued*

Shot number	Charge of explosive (lb.)	Shot hole depth (ft.)	Shot location	Recording location	Distance X (ft.)	Travel Time			
						U (sec.)	S (sec.)	C (sec.)	W (sec.)
Station 2, October 18, 1935									
9	2	.....	1	A	2540	.450	.....	.....	.514
10	4	.....	2	A	4920	.917	.....	.....	.991
11	4	.....	3	A	7060	1.290	1.192	.....	1.426
12	6	.....	4	A	8600	1.583	1.332	.....	1.738
Mathews, July 28–31, 1935									SH
2		18	1	9	534	.....	.....	.....	.088
7/31				7	1112	.207	.....	.....	.146
				5	1655	.316	.....	.....	.260
				4a	2021	.386	.....	.....	.347
1	6.5	18	1	4a	2021	.386	.....	.....	.352
7/31				1	2975	.576	.....	.....	.....
				15	3993	.787	.....	.....	.....
				17a	4925	.986	.....	.929	.....
1	8.5	18	1	4000	5737	1.133	.....	.962	.....
7/30				3000	6706	.....	1.238	1.012	.....
				2000	7686	.....	1.385	1.063	.....
				1000	8664	.....	.....	1.114	.....
1	.....	18	1	4000	5737	1.125	.....	.963	.....
7/29				3250	6498	1.264	.....	.987	.....
				2750	7018	.....	1.250	1.030	.....
				2000	7686	.....	1.382	1.067	.....
1	6.5	18	1	2000	7686	.....	1.341	1.057	.....
7/28				1000	8664	.....	1.486	1.119	.....
				0	9650	.....	1.635	1.157	.....

membered that the results given for the two most easterly stations are based on meager data and on the assumption that the velocities are roughly the same as at neighboring stations.

REFRACTION MEASUREMENTS ON THE WOODS HOLE SECTION

LOCATION

The Woods Hole section lies along the meridian 70°43′ West. The position of the stations along this line are measured in statute miles south of the 42nd Parallel. Four stations were occupied for these, the first refraction, measurements made at sea. The primary purpose of these measurements was to develop the technique and to make certain that the instruments were working properly before getting too far away from the base at Woods Hole, Massachusetts. Although the results are incomplete, they offer enough substantiation to those obtained at Cape Henry to justify inclusion here.

TABLE 3.—*Calculations and results for refraction measurements*

Station	Vw ft./sec.	Vu ft./sec.	Tu sec.	hw ft.	Vs ft./sec.	Ts sec.	hu ft.	Vc ft./sec.	Tc sec.	hs ft.	Elev. Surf. ft.	Elev. semicon. ft.	Elev. cryst. ft.
Camp Lee	.....	6381	.....	0	.....	.....	342.6	12290	.0918	0	+110	.....	− 230
Youngbloods store	.....	5532	.....	0	.....	.....	401.0	12850	.1309	0	+140	.....	− 260
Prince George	2000*	5999	.0210	22.3	.....	.....	403.5	20830	.1510	0	+130	.....	− 300
Newville	2000*	5741	.0383	40.9	.....	.....	377.0	18340	.1655	0	+120	.....	− 300
Brandon	2000*	5145	.0101	11.0	.....	.....	361.0	17140	.1450	0	+ 70	.....	− 300
Mercy Seat Church	2000*	4894	.0010	1.1	.....	.....	929	15180	.3607	0	+100	.....	− 830
Surry 500	.....	5605	.....	0	.....	.....	1164	18790	.3965	0	+ 90	.....	−1070
Surry 9500	.....	5660	.....	0	.....	.....	1152	18820	.3951	0	+ 90	.....	−1060
Mulberry Island	.....	5376	.....	0	6331	.1086	553	19430	.4223	752	+ 10	− 540	−1300
Hampton	.....	5645	.....	0	6714	.1102	575	20170	.5761	1353	+ 10	− 570	−1920
Fort Monroe	.....	5296	.....	0	6800*	.251	1060	19000*	.713	1198	0	−1060	−2260
Cape Henry	2000*	5551	.0207	22.2	7347	.3949	1582	17470	.8841	1303	+ 10	−1590	−2900
Station 8	.....	5644	.....	0	7640	.4194	1756	17200	1.160	2439	− 50	−1810	−4250
Station 12	.....	5650	.....	0	7435	.2513	1092	18920	1.703	5410	− 80	−1170	−6580
Station 16	.....	5595	.....	0	8180*	.2012	772	18000*	>2.56	>10600	−130	− 900	<−11500
Station 20	.....	5635	.....	0	8390*	.1254	475	18000*	2.65	11810	−210	− 680	−12500
Woods Hole	.....	5360	.....	0	.....	.....	235	17840	.084	0	− 60	.....	− 295
Station 1	.....	5670	.....	0	.....	.....	1227	17800*	.410	0	−110	.....	−1337
Station 3	.....	5640	.....	0	13460	.519	1607	17800*	>.720	>1843	−170	−1780	<−3620
Station 2	.....	5440	.....	0	11000	.550	1722	17800*	>.849	>1725	−200	−1920	<−3650
Mathews	.....	5099	.....	0	7294	.309	1102	19725	.672	1001	+ 15	−1087	−2088

* Values taken either by assumption or by interpolation from Figure 31.

Table 4 shows the stations of the Woods Hole section, together with the position and depth of water for each. The method of procedure used here was identical with that described for the preceding section.

TABLE 4.—*Refraction stations, Woods Hole section*

Station	North Latitude	Statute miles south of 42nd Parallel	Depth of water (ft.)
Woods Hole	41°31.2′	33.2	60
Station 1	41°09.8′	57.8	110
Station 3	40°57.2′	72.3	170
Station 2	40°43.4′	88.2	200

STATIONS

Woods Hole.—Figure 33 shows the travel-time curves for all stations of the Woods Hole section. The first measurement was made while the ship, *Atlantis,* on which the recording instruments were installed, was at dock in Woods Hole. It is a good measurement, which yields a normal velocity of 17,840 ft/sec for the crystalline rocks. The water wave was notably weak on all shots at this station, probably because the water in the harbor was shallow, and the bottom was irregular.

No. 1.—The measurement for No. 1 station is incomplete. The U branch of the travel-time curve is well determined, but there is available only a single point for determining the remaining branches of the curve. The data do not suffice to determine whether the first arrival on the longest shot is an S or a C phase, so calculations are made on both assumptions. By assuming it to be an S phase, and taking the velocity in S to be 11,000 ft/sec, according to interpolation from Figure 31, the thickness of the unconsolidated zone is found to be 960 feet. On this assumption, a crystalline zone, with velocity 17,800 ft/sec, would have to be at least 1830 feet below sea level in order to escape detection by the longest shot of the station.

On the other hand, if the first arrival on the longest shot is assumed to be a C phase, whose velocity is considered to be 17800 ft/sec, by interpolation in Figure 31, the depth to the crystalline rocks is 1230 feet. This last result is accepted as the proper interpretation of the data, because it appears to be more concordant with the facts and probabilities, as shown in Figure 32.

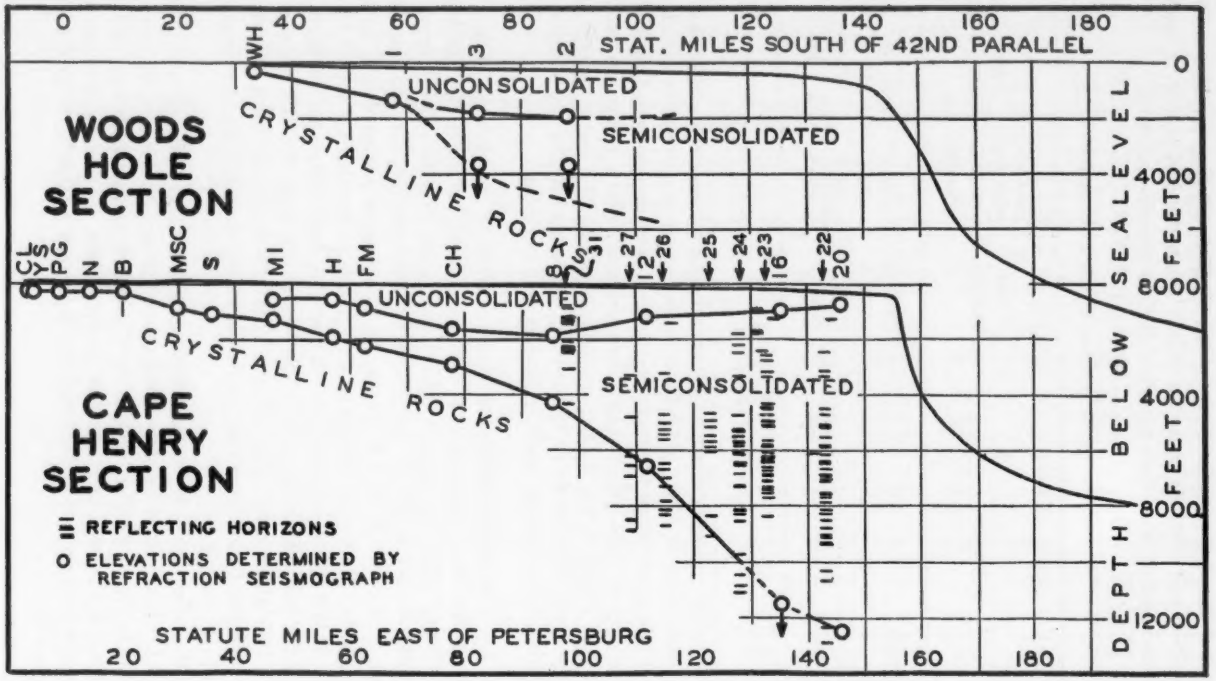


FIGURE 32.—*Cape Henry and Woods Hole sections*

No. 3.—The measurement at No. 3 station is incomplete. None of the velocities is well determined. The observed velocity of 13,400 ft/sec is considered to be the S phase and too large, because of experimental error. On this interpretation, it is necessary that the crystalline zone be at least 3600 feet below sea level in order to escape detection by the longest shot of the station.

No. 2.—The U phase is well determined at No. 2 station. The observed velocity 11,000 ft/sec is considered to represent the S phase, despite the fact that this velocity is much higher than that found for S on the Cape Henry section. Presumably, the crystalline rocks, whose velocity is assumed to be 17,800 ft/sec, are too deep to be detected by the longest of the shots. In that case, they must be at least 3600 feet below sea level.

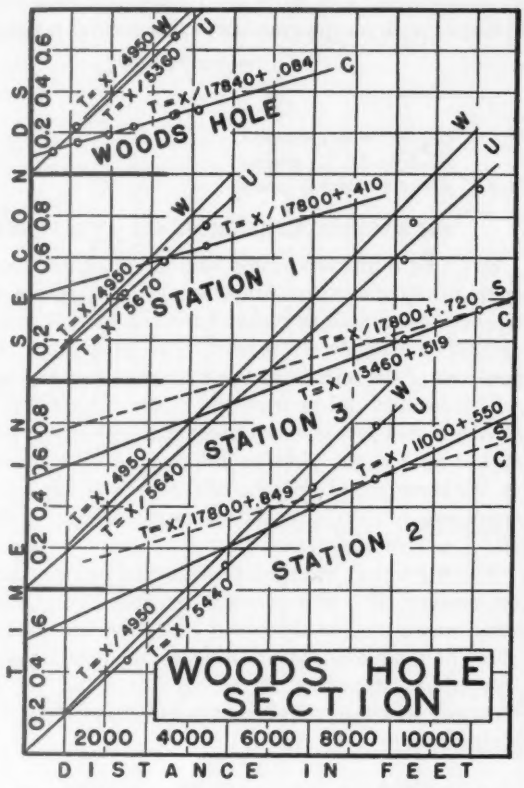


FIGURE 33.—*Travel-time curves for Woods Hole section*

SUMMARY

Tables 2 and 3 contain the travel-time data, the calculations, and the results for the Woods Hole section. The results are shown graphically in Figure 32. Although these results are incomplete, it is possible to make a useful comparison with those from Cape Henry in regard to the thickness of the unconsolidated zone. The two sections, as arranged in Figure 32 for ready comparison of points at equal distances from the outer edge

of the shelf, show that all results obtained at Woods Hole are in good agreement with those from Cape Henry.

The velocity 17,800 ft/sec for crystalline rocks, which was obtained at Woods Hole, is in fairly good agreement with values found by Ewing in earlier work on the granites of this general region: [9]

	Velocity of longitudinal waves ft/sec
Quincy (Mass.) granite	16260
Westerly (R. I.) granite	16400
Rockport (Mass.) granite	17800

REFRACTION MEASUREMENTS AT MATHEWS, VIRGINIA

One long refraction profile and several shorter auxiliary profiles were made in the vicinity of Mathews, Virginia, where a deep bore-hole had reached the crystalline rocks at a depth of 2307 feet beneath the surface, or 2297 feet below sea level.[10] The location of the bore-hole and the positions of the seismic stations are shown in Figure 34.

This locality differs from the others discussed in the present work in having a thin bed of moderately high velocity material (8000 to 10,000 ft/sec) at a depth of 70 to 100 feet. This bed causes the first arrival on the three points between 1000 and 2000 feet on the main travel-time curve and is clearly displayed on the auxiliary curves for locations H and I, as shown in Figure 35.

The wave that travelled through this bed is entered in Table 2 under the heading SH. It was easily recognizable on the seismograms, by its small amplitude, its sinusoidal character and gradual beginning, and its habit of gradually disappearing for long distances. Its behavior with increasing distance led to the conclusion that the bed through which it travelled was thin. The depths calculated for this bed at the various stations are:

Station	V_1 (ft/sec)	V_2 (ft/sec)	T (sec)	h (ft)
Mathews	5099	10000	0.034	101
Location I	4386	9091	0.027	68
Location H	4300	7950	0.028	72

The log of the well reports iron-cemented gravel at 83 to 200 feet depths. It is reasonable to associate the bed detected by the seismic measurements with this horizon. This bed is considered to be so thin that it may be neglected altogether in calculation of depths to underlying horizons.

[9] L. D. Leet and M. Ewing: *Velocity of elastic waves in granite*, Physics, vol. 2 (1932) p. 160.

[10] W. C. Mansfield: *An oil-prospecting well near Havelock, North Carolina*, N. C. Dept. Conserv. Devel., Econ. Pap. no. 58 (1927) 19 pages.

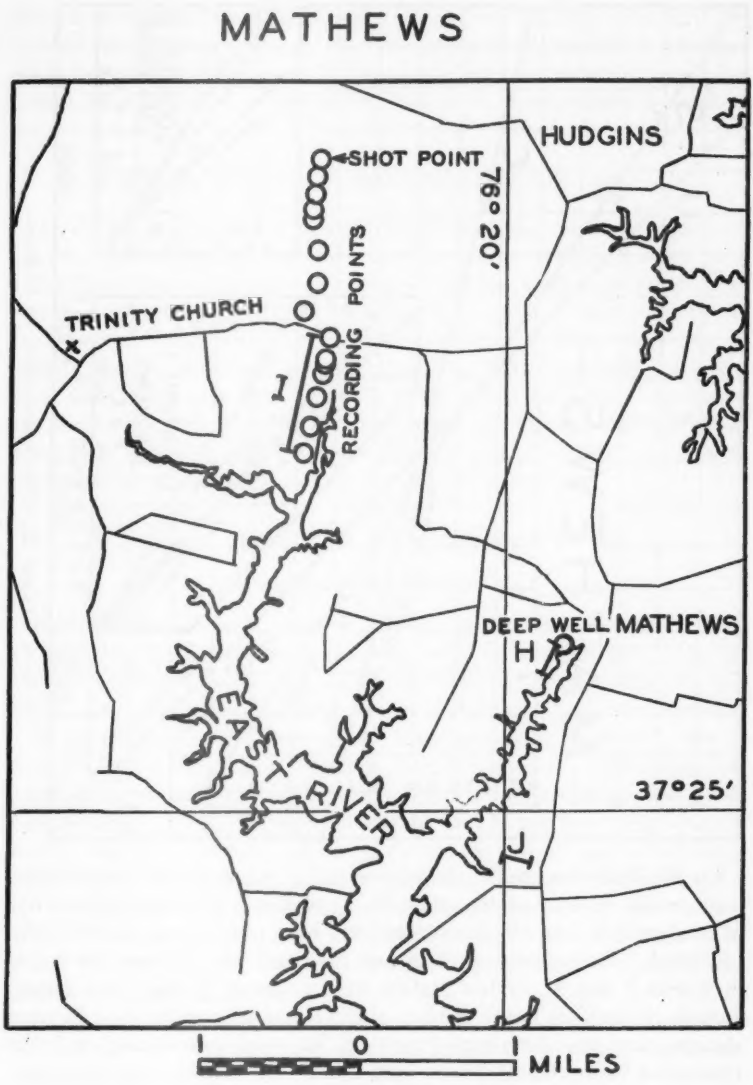


FIGURE 34.—*Station map for Mathews*

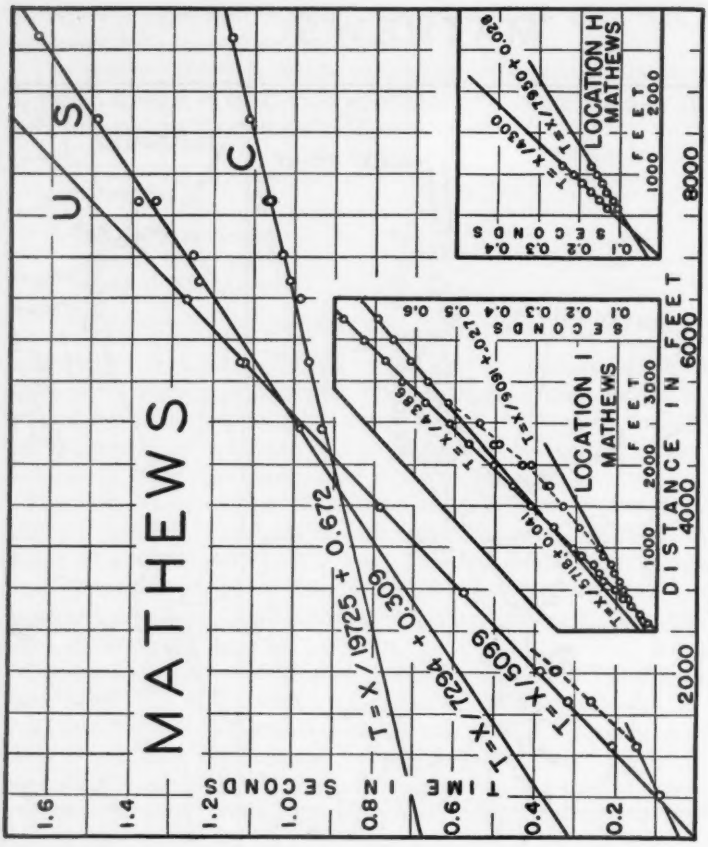


FIGURE 35.—*Travel-time curves for Mathews*

On the Mathews profile, the phases U, C, and S are present, as they were on the stations of the Cape Henry section. The seismograms are of good quality, and all phases that have been plotted may be definitely identified. The travel-time data and the depth calculations are shown in Tables 2 and 3. If this station were projected to the Cape Henry section, it would lie in the vicinity of the Hampton and the Fort Monroe stations, and the elevations found for the semi-consolidated and the crystalline zones would be in agreement. To compare the elevation —2088 feet, found for the crystallines at the Mathews Station, with the —2297 feet, found in the bore-hole at Mathews, it is necessary to make allowance for the fact that the station is about 2½ miles from the well. Assuming the regional slope in the surface of the crystalline rocks to be

the same as that found in the vicinity of Fort Monroe (Fig. 32), the change in elevation between the two points is estimated at 140 feet. This reduces the discrepancy in the seismic result to about 70 feet. The discrepancy would have been large, had the S phase, which is masked, been overlooked.

SEISMIC REFLECTION MEASUREMENTS ON LAND

At Hampton, and also at Mathews, attempts were made to obtain reflections from the surface of the crystalline rocks. Although a few poor reflections were recorded from points within the semi-consolidated material at Hampton, these tests were almost completely unsuccessful. It was not practical, with the facilities available, to bury the explosive deeper than 15 or 20 feet, so the failure may have been caused by surface disturbance and "ground roll", which would have been eliminated by deeper shot holes, similar to those commonly used in land work. It also seems probable that the crystalline rocks in this region do not present an efficient reflecting surface for seismic waves.

SEISMIC REFLECTION MEASUREMENTS AT SEA

METHOD OF OPERATION

Seismic reflection tests at sea were made at the stations marked 22 to 39 in Figure 2. The *Oceanographer* was used for stations 26A, 27A, 28-39, and the Atlantis for stations 22-27. The apparatus used was the same as that described in connection with the refraction measurements.

In all reflection tests made at sea, three geophones were used, one at the bow of the ship, one at the stern, and one midship. The geophones were connected through transformer-coupled amplifiers to the three galvanometers in one oscillograph, which was used to produce a high-magnification record. The other oscillograph was used to produce a low-magnification record by shunting all its galvanometers and connecting each of them in series with the corresponding galvanometer of the first oscillograph. The use of two degrees of magnification doubled the range over which the amplitude of recorded motion was suitable for showing the presence of reflected waves.

In making a test, the three geophones were simultaneously lowered to the bottom from their positions along the length of the ship. The bomb was placed on bottom, at a distance of 100 to 1000 feet astern. As soon as the geophones were on bottom the bomb was exploded electrically by means of a circuit controlled by the operator of the recording apparatus. The firing current was adjusted to require about one-tenth second for producing detonation, and the instant at which the circuit through the deto-

nator was broken was recorded on the seismogram and taken as the instant of the explosion. Bombs for holding T. N. T. were made from 6 by 1½-inch iron pipe nipples, closed at one end by a cap and at the other by a reducer and packing gland, through which the leads to the

TABLE 5.—*Reflection stations at sea*

Station	Statute miles east of Petersburg	Depth of water (ft.)	Time in unconsolidated zone (sec.)*
26A	131	108	0.30
27A	130	120	0.30
28	131	126	0.30
29	151	282	0.12
30	142	204	0.21
31	98	66	0.59
32	116	90	0.37
33	132	144	0.29
34	142	186	0.21
35	147	246	0.15
36	134	138	0.28
37	153	312	0.10
38	149	246	0.14
39	72	36	0.50
22	143	162	0.20
23	133	120	0.29
24	128	90	0.31
25	123	84	0.34
26	115	84	0.37
27	109	72	0.43

* These values computed from the refraction results shown in Figure 32.

detonator were introduced. In the later experiments, a waterproof gelatin explosive was used, which eliminated the necessity for waterproof bombs. In the first tests the bomb was thrown overboard while the ship was being brought to a stop, the three geophones were dropped as soon as the ship was completely stopped, and the test was made within two or three minutes thereafter.

Captain N. H. Heck devised a method of making reflection tests while the ship was anchored. The bomb was attached to a float and lowered from the stern of the ship.[11] When the tide had carried the float to the desired distance astern, a trip released the bomb and permitted it to drop

to the bottom. The geophones were then lowered, and the test carried out as already described. This method was of great value, because it permitted repetitions at each station until the proper charge of explosive and the proper degree of amplification for best results had been determined.

In the reflection work done on the *Atlantis,* the bombs were carried astern, while the ship was at anchor, by a crew of men in a whaleboat. The whaleboat remained at its station until all necessary repetitions had been made. This method was by far the most satisfactory of those tried, but it would have been impractical to use it in rough weather.

METHOD OF INTERPRETATION

A wave returning to the surface as an echo from a horizontal reflecting surface at considerable depth is identified on a seismogram by the fact that its travel-time is practically the same as that for geophones set at varying distances from the shot point. The total spread of the geophones was 110 feet on the *Atlantis* and almost 200 feet on the *Oceanographer,* and the shot was placed at distances of 100 to 1000 feet from the nearest geophone. Within the limit of accuracy of the present work, it is permissible to consider, in making calculations of depth, that the reflected waves travel vertical paths.

The velocities in the two uppermost zones and the thickness of the upper one are considered to have the values given for them by the refraction measurements, as recorded in Figures 31 and 32. The elevation of a reflecting surface in the semi-consolidated zone is calculated from the equation,

$$E = h_w + h_u + \tfrac{1}{2}\left(T - \frac{2h_u}{V_u}\right)V_s$$

where

E = depth (below sea level) of reflecting surface
h_w = depth of water
h_u = thickness of unconsolidated zone
T = travel-time of reflected wave
V_u = velocity in unconsolidated
V_s = velocity in semiconsolidated
$2h_u/V_u$ = time in unconsolidated.

RESULTS

The stations at which reflection measurements were made are shown in Figure 2 and are listed, with accompanying data, in Table 5. Each station is projected to the line of the Cape Henry section, and the dis-

[11] N. H. Heck: *Sounding to granite,* U. S. Coast Geodet. Surv., Field Eng. Bull., vol. 9 (1935) p. 62.

TABLE 6.—*Reflection tests at sea*

Shot No.	Station No.	Distance (ft.)	Charge (lb.)	Amplifier (stages)
50	26A	.....	0.5	3
51	"	.....	"	2
52	27A	.....	"	"
53	28	.....	"	"
54	"	.....	"	"
55	29	.....	"	"
56	30	.....	"	"
57	31	.....	"	"
58	"	.....	"	"
59	"	.....	det	3
60	"	.....	"	"
61	"	.....	0.5	"
62	32	.....	"	2
63	33	.....	"	"
64	34	.....	"	"
65	35	.....	0.2	"
66	36	.....	0.5	3
67	"	.....	det	"
68	"	.....	"	"
69	"	.....	0.3	2
70	37	.....	0.5	"
71	38	.....	"	"
72	39	.....	det	3
36	22	500	0.5	"
37	"	550	0.3	"
38	"	600	0.3	"
39	"	500	0.5	"
40	"	"	"	"
41	"	1050	2.0	"
42	"	"	"	"
43	23	850	"	"
44	"	900	"	"
45	24	800	"	"
46	"	750	10.0	"
47	25	700	2.0	"
48	26	600	"	"
49	27	700	"	"

tance of the projected point from Petersburg is taken as the coördinate for the station.

Table 6 shows the data on charge of explosive and degree of amplification for each test that was made. The explosive used was T. N. T., 60 per cent dynamite, or, in some cases, only a detonator.

TABLE 7.—*Depths measured by reflections*

Reflection time (sec.)	Elevation reflector (ft.)
(Shot 53)	
0.28	920
0.38	1690
(Shot 54)	
0.39	1730
(Shot 57)	
0.41	1220
0.48	1420
0.68	2050
0.80	2500
(Shot 58)	
0.29	860
0.42	1250
0.48	1420
(Shot 59)	
0.42	1250
(Shot 60)	
0.41	1220
0.45	1340
(Shot 61)	
0.48	1420
0.79	2460
0.96	3100
1.30	4360
(Shot 62)	
0.44	1400
(Shot 63)	
0.66	2400
(Shot 64)	
1.58	6120
(Shot 66)	
1.11	1250
(Shot 38)	
1.47	6010
(Shot 39)	
0.61	2430
1.43	5850
1.51	6180
1.57	6430
1.63	6680
1.74	7140
1.84	7560
2.14	8800
2.23	9170
(Shot 40)	
1.10	2460
1.16	4720
1.26	5130
1.57	6430
1.70	6970
1.85	7600
1.92	7890
2.27	9390
(Shot 41)	
1.26	5130
1.44	5890
1.58	6470
1.72	7050
1.87	7680
2.01	8270
(Shot 42)	
0.78	3140
0.84	3390
0.90	3640
1.26	5130
1.36	5560
1.70	6970
1.98	8140
2.07	8520
2.20	9050
2.28	9390
2.49	10270
2.59	10670
2.99	12340
3.10	12920
(Shot 43)	
0.70	2580
0.74	2750
0.80	3360
1.17	4500
1.28	4940
1.34	5180
1.49	5800
1.55	6030
1.63	6360
1.75	6840
1.82	7140
(Shot 44)	
0.84	3160
0.93	3520
0.96	3640
1.03	3920
1.12	4300
1.16	4460
1.22	4700
1.28	4940
1.34	5180
1.50	5840
1.60	6240
1.65	6440
1.89	7420
1.93	7580
1.97	7750
2.13	8400
(Shot 45)	
0.52	1770
0.62	2200
0.69	2480
0.90	3320
1.40	5340
1.51	5770
1.59	6090
1.71	6570
1.88	7260
2.14	8300
2.21	8530
2.51	9790
2.68	10460
(Shot 46)	
1.26	4760
1.39	5290
1.44	5490
1.53	5850
1.69	6490
1.80	6850
1.89	7300
2.09	8100
2.78	10860
2.85	11160
(Shot 47)	
1.07	3910
1.26	4660
1.33	4940
1.39	5180
1.47	5490
1.52	5690
1.60	6010
2.20	8380
2.38	9100
(Shot 48)	
0.92	3270
0.97	3470
1.30	4750
1.37	5020
1.44	5290
1.48	5440
1.53	5640
1.75	6500
1.81	6730
1.88	7000
1.96	7310
2.10	7860
2.18	8180
2.23	8370
2.33	8760
(Shot 49)	
0.97	3340
1.04	3620
1.09	3800
1.35	4800
1.72	6210
1.81	6560
1.91	6930
2.32	8500
2.41	8850

Table 7 shows the travel-times of the reflected waves that were recorded in each test and the depths of the reflecting surfaces which were calculated from them. The position of these surfaces is indicated by short horizontal marks in Figure 32.

CONCLUSIONS

A striking feature of the reflection results is the great number of reflecting surfaces revealed. Some of the identifications of phases on the seismograms as reflections may be open to question, but many of them are uncommonly good. It is concluded that there are numerous

horizons in the semi-consolidated zone at which slight variations in composition or lithification of the sediments are sufficiently abrupt to produce measurable reflections of the elastic waves. If several tests were made at a single station, it is the general rule that the reflections that look best on one record are also prominent on the others, but there are disappointing exceptions to the rule. The object of the repetitions was to find a combination of operating conditions that would cause one reflection—that from the surface of the crystalline rocks—to stand out with such prominence that it could be identified unmistakably, and, consequently, variations were introduced intentionally into the conditions of each test at a given station (*see* Table 6). As no reflection was found that could be identified with the surface of the crystalline, it will probably not be possible to map the crystalline in this region by reflections.

Some tentative correlations of the more prominent reflections from one station to the next have been attempted, but the stations are so far apart that no result reliable enough for publication has been reached.

To account for the ease with which reflections were observed at sea, in contrast to the difficulty experienced in detecting them on land, it is suggested that the water and the sediments of the unconsolidated zone are sufficiently similar in elastic properties to behave as a single medium. Then, at sea stations, the bomb and, also, the geophones are far enough away from the free surface of the medium to escape disturbance from surface effects. The data on velocities and the absence of "ground roll" in refraction measurements at sea support this view.

The reflections obtained from points within the semi-consolidated zone are good enough so that structure within the semi-consolidated zone could be studied and mapped in this region by the reflection method.

From Figure 32, it is seen that the reflection results give good qualitative support for the section constructed from refraction measurements. At stations 22-25 and 31, reflections are obtained to depths approximately equal to those given for the crystalline rocks. At stations 26 and 27 the reflections have been picked from points considerably beneath the indicated position of the crystalline surface. The deeper reflections for these stations are somewhat questionable in appearance. In general, the part of the section from which reflections are reported coincides reasonably well with the part which had been classified as semi-consolidated from the refraction results.

SUMMARY

The refraction seismograph and the reflection seismograph were successfully adapted for work at sea in depths up to 100 fathoms.

Refraction measurements on land and at sea near Cape Henry, Vir-

ginia, and Woods Hole, Massachusetts, determined the configuration of the surface of the basement crystalline rocks, from the foot of the Piedmont Plateau to the edge of the Continental Shelf (Fig. 32). The thickness of unconsolidated and semi-consolidated material (velocity of longitudinal waves about 8000 ft/sec) near the edge of the Continental Shelf was found to be about 12,000 feet.

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GEOPHYSICAL INVESTIGATIONS IN THE EMERGED AND SUBMERGED ATLANTIC COASTAL PLAIN

PART II: GEOLOGICAL SIGNIFICANCE OF THE GEOPHYSICAL DATA

BY BENJAMIN L. MILLER

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INTRODUCTION

The use of geophysical methods for the solution of geological problems has been rapidly gaining favor during the past few years, largely due to the great success attained by those engaged in petroleum exploration. It is particularly fortunate that geophysics should have early commended itself to commercial concerns, for, otherwise, lack of funds might have hampered its progress. Geologists who are engaged in non-commercial investigations have appreciated the importance of this new tool but have been prevented from using it extensively because of the relatively great expense involved as well as the difficulty of securing properly trained men to collect the geophysical data. This paper is concerned with the geological interpretation of results obtained in a geophysical investigation of somewhat novel character, under the leadership of Maurice Ewing. It accompanies a paper by Ewing, Crary, and Rutherford, which should be read in conjunction with this article, in order properly to comprehend the project and results.

STATUS OF OCEAN-BOTTOM INVESTIGATIONS

A brief survey of existing information concerning the rocks forming ocean basins seems appropriate. The investigations thus far carried on

have been concerned mainly with the determination of the topography, the character of the deposits now forming on the ocean bottoms, and the density of the rocks.

It seems natural that the topography of the ocean basins should receive attention first. Sounding with weight and line had yielded a great amount of information before the advent of sonic sounding. Since the development of the sonic method, knowledge of the configuration of the ocean

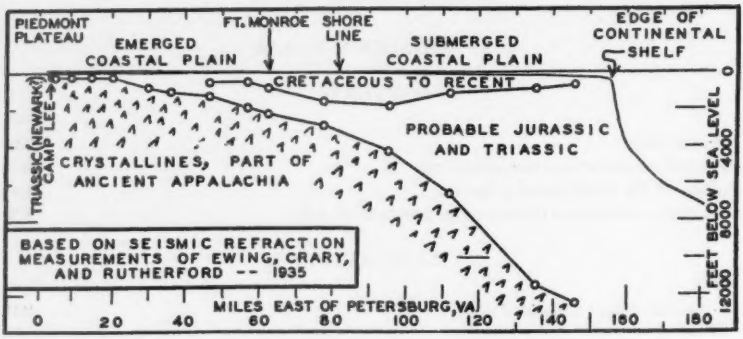
PIEDMONT PLATEAU
EMERGED COASTAL PLAIN
FT. MONROE SHORE LINE
SUBMERGED COASTAL PLAIN
EDGE OF CONTINENTAL SHELF
CRETACEOUS TO RECENT
PROBABLE JURASSIC AND TRIASSIC
TRIASSIC (NEWARK) CAMP LEE
CRYSTALLINES, PART OF ANCIENT APPALACHIA
BASED ON SEISMIC REFRACTION MEASUREMENTS OF EWING, CRARY, AND RUTHERFORD -- 1935
MILES EAST OF PETERSBURG, VA
0 20 40 60 80 100 120 140 160 180
FEET BELOW SEA LEVEL
0 4000 8000 12000

FIGURE 1.—*Section across Coastal Plain*

From the Piedmont Plateau to the outer edge of the Continental Shelf.

bottoms has made rapid advance, until one visualizes the eventual preparation of contour maps of the oceans that will compare favorably with our land topographic maps.

Ever since the Challenger Expedition, there has been great interest shown in the investigation of the deposits now forming on the ocean bottom, and enough data are now at hand to permit one to classify the materials and roughly to map their distribution over rather extensive areas. Although information of this kind is useful, there has been a great desire to know what lies beneath the immediate surface. A recent apparatus [1] has been devised by which cores up to 8⅔ feet long, and from depths of 7500 feet, have been obtained. Dredging [2] along the walls of the submerged canyons cut in the margins of the Continental Platform has shown Tertiary and Cretaceous strata lying beneath Recent deposits. However, dredging in some of the submerged canyons has shown only Recent forms. At best, cores such as have been obtained up to the present, and by dredging in the canyons, furnish only limited information of the char-

[1] C. S. Piggott: *Apparatus to secure core samples from the ocean-bottom*, Geol. Soc. Am., Bull., vol. 47 (1936) p. 675-684.

[2] H. C. Stetson, L. W. Stephenson, R. S. Bassler, J. A. Cushman: *Geology and paleontology of the Georges Bank Canyons*, Geol. Soc. Am., Bull., vol. 47 (1936) p. 339-440.

acter of the rocks of the ocean bottom. Deep borings in the ocean bottom have been discussed, but no one has yet devised any satisfactory method, although by some persons it is considered feasible.

Variations in density of the rocks of the ocean floor have been determined in various places throughout the world. In this field, Dr. F. A. Vening Meinesz has been the leader. The most recent published work[8] of this kind in the western Atlantic Ocean was done in a government submarine. The report of another similar gravity investigation, carried on by the United States Navy and Maurice Ewing, Lieut. Hoskinson, and H. H. Hess, during December 1936 and January 1937, will be published shortly.

More generalized information concerning the varying densities of the sub-oceanic rocks has been obtained by determination of the different rates of earthquake wave propagation across the oceans.

Summarizing, one may say that at the present time, there is fairly accurate information concerning the major topographic characteristics of the ocean basins and considerable knowledge concerning the character and distribution of the sediments now accumulating in the ocean basins, and, finally, there is evidence that the rocks vary in density, indicating different types of materials.

All these lines of investigation have yielded much valuable information, but they lack much desired data. In the absence of direct evidence, geologists have long theorized about the rocks concealed by the ocean waters. The literature contains frequent references to land bridges, submerged land masses, both large and small, and permanence of ocean basins. The former existence of Atlantis is believed by some to be as authentic as that of Appalachia. By many, the rocks of the ocean bottoms are believed to hold the key to many stratigraphic problems and particularly to the correct interpretation of geologic events that concern both the continents and the oceans. Geologists would like to know the character, thickness, age, structure, and relationships of the rocks of the ocean basins, downward for several thousand feet. The rocks of three-fourths of the surface of the earth still remain practically unknown. Under these conditions, any method or process by which geologists can obtain any information concerning the sub-oceanic rocks is particularly welcomed.

Some statements by de Launay in 1913 still express well the status of ocean-floor geology and the wish of many geologists for more information. He says:

"The geology of the bottom of the seas is still very poorly known to us. . . . It is the almost unexplored field where one may hope some day to find the key to prob-

[8] R. M. Field, T. T. Brown, F. A. Vening Meinesz, and H. H. Hess: *The Navy-Princeton Gravity Expedition to the West Indies in 1932*, U. S. Hydrographic Office (1933) 54 pages.

lems that terrestrial geology elsewhere proposes to us. . . . It may be hoped that some day, instead of limiting the work to the epidermis of these submarine deposits, we shall penetrate them better by deeper and still deeper borings."[4]

As man has not yet succeeded in making these oceanic borings, other means for determining the character of the rocks of the ocean floor have been sought. Geophysical methods are borrowed from the oil industry, for the first time for this purpose. This paper is concerned with the application of seismic geophysical work only. Whether or not, some of the other geophysical methods may eventually be used is problematical.

OUTLINE OF THE GEOPHYSICAL PROJECT

The plan to conduct seismic geophysical investigations across the Coastal Plain, both in the emerged and in the submerged portions, was formulated during the winter of 1934-1935, in a series of conferences with officials of the United States Coast and Geodetic Survey and the American Geophysical Union, and the project later discussed with officials of the United States Geological Survey, the Geological Society of America, and the Woods Hole Oceanographic Institution. Particular credit in perfecting the plan is due to Dr. R. M. Field, of Princeton University, and to Dr. William Bowie and Dr. N. H. Heck, of the United States Coast and Geodetic Survey. In May 1935, a grant by the Geological Society of America from the Penrose Bequest, to B. L. Miller and Maurice Ewing, for the purchase of necessary equipment to supplement that possessed by Lehigh University and for the maintenance of an investigating party, made it possible to undertake the project.

The purpose of the investigation was to secure exact information in regard to the basement of the Coastal Plain sediments, both in the emerged and in the submerged portions, from the inner margin at the border of the Piedmont Plateau to the edge of the Continental Slope, to determine the thickness of the Coastal Plain sediments and, if possible, some facts concerning their character. The importance of such information is readily apparent, especially to the stratigrapher and the student of oceanographic problems.

The place selected for the investigations was along a line across the Coastal Plain of Virginia, from Petersburg to Cape Henry, and thence eastward across the Continental Shelf to the inner edge of the Continental Slope. The selection was primarily due to the fact that the *Oceanographer*, of the United States Coast and Geodetic Survey, had been assigned to the area lying to the east of the Virginia capes for the summer's work. The line chosen was particularly favorable, because reliable records of two deep wells, drilled to the underlying crystalline rocks near the mouth

[4] L. de Launay: *The geology of the bottom of the seas*, Smithson. Rept. 1914, Publ. 2334 (1915) p. 329-352.

of Chesapeake Bay, were available. This permitted Ewing and his associates to determine rates of vibration propagation in known rocks before entering the entirely unknown region. In this way, a valuable check was possible.

Inasmuch as this paper is based on the results described in the accompanying contribution by Ewing, Crary, and Rutherford, it is unnecessary to repeat the descriptions. The geophysicist employing the seismic method furnishes the geologist with data concerning the relative speeds of the seismic vibrations and leaves to the geologist the explanation of the results. The writer, therefore, offers some possible interpretations. It should be clearly recognized that these are tentative, as they are based on limited facts and, therefore, may need to be modified or discarded when more complete information is obtained. Everyone connected with the project is convinced that the methods employed are reliable and the results dependable. Likewise, all agree that additional geophysical investigations should be carried on along similar lines, across the Coastal Plain in many more sections. They should also be extended eastward at least to the foot of the Continental Slope and preferably much farther. It is believed that this can, and will, be done eventually and, when done, will furnish much valuable information for interpreting the geologic histories of the continents as well as that of the oceans. If one had as much information about the rocks of the ocean bottoms as is available about the rocks of the continents, probably many accepted theories would be materially modified. The work of Ewing and his associates seems to point the way to the enlargement of knowledge of the ocean basin rocks.

GEOLOGICAL CONCLUSIONS BASED ON THE GEOPHYSICAL DATA

The geophysical data, described in detail by Dr. Ewing and his associates, indicate three different kinds of rocks composing the submerged portion of the Coastal Plain, in which the seismic vibrations have different speeds. The deepest transmit the vibrations at the rate of 17000 to 20000 feet per second; above this is a zone of rocks in which the speed is 6500 to 8500 feet per second; an upper zone has speeds of 5000 to 6000 feet per second. The following tentative geologic explanations are proposed. Reference should be made to the figures of the article by Ewing, Crary, and Rutherford.

(1) There can be little doubt that the rocks with the greatest vibration speed are crystalline rocks of the type well exposed in the Piedmont Plateau. Near the inner margin of the Coastal Plain, in the vicinity of the Fall Line, the unconsolidated Coastal Plain sediments overlie a complex of igneous and metamorphic rocks, composed of granites, gneisses, and schists. Some deep borings for water or petroleum at several places in the

Coastal Plain of New Jersey, Virginia, and North Carolina near the ocean, and, hence, considerable distances from the surface exposure of the Piedmont metamorphic rocks, have encountered similar rocks. The cuttings show feldspar, quartz, mica, and all have been roughly termed "granite".[5] These are summarized in the following table:

TABLE 1.—*Borings in the Atlantic Coastal Plain extending to the basal crystalline rocks*

Location	Depth below sea level (feet)
Jackson's Mills, New Jersey	1226
Mathews, Virginia (three-quarters of a mile southwest of Mathews Court House)	2300
Fort Monroe, Virginia	2240
Wilmington, North Carolina	1109
Fort Caswell, North Carolina (mouth of Cape Fear River)	1540
Havelock, North Carolina (19 miles northwest of Beaufort)	2318

In this table, the Jackson's Mills well lies some distance inland, at an elevation 110 feet A.T. The crystalline floor was struck at a depth of 1336 feet. The other borings were on low ground in proximity to the ocean. In some of the wells the exact depth to the crystalline floor is somewhat doubtful, but the figures are probably accurate within a few feet. Ewing used the depths of the first three borings in determining speed of vibration propagation in the strata penetrated in these wells.

The seismic results furnish corroborative evidence for the belief that dense basal rocks underlie both the emerged and the submerged portions of the Atlantic Coastal Plain. The speed of propagation of the vibrations in the rocks lying beneath the Cretaceous sediments in the emerged Coastal Plain was essentially the same as that in the basement rocks of the submerged portion, thus leading to the conclusion that the basement rocks, from the inner edge of the Coastal Plain to the outer edge of the Continental Shelf, are probably all metamorphosed crystalline rocks. The deep borings listed indicate that the basal crystalline rock floor everywhere slopes seaward but not at the same rate. In New Jersey, the Jackson's Mills boring indicates a slope of approximately 65 feet to the mile; the two Virginia borings indicate a slope of 30 feet per mile; and the North

[5] M. E. Johnson: *The mineral industry of New Jersey for 1929*, N. J. Dept. Conserv. Devel., Geol. Ser., Bull. 30 (1931) p. 28.

Samuel Sanford: *The underground water resources of the Coastal Plain province of Virginia*, Va. Geol. Surv., Bull. 5 (1913) p. 100-101.

L. W. Stephenson: *The Coastal Plain of North Carolina; the Cretaceous, Lafayette, and Quaternary formations*, N. C. Geol. Econ. Surv., vol. 3 (1912) p. 163-166; (with B. L. Johnson) *Water resources of the Coastal Plain of North Carolina, ibid.*, p. 442-443.

L. W. Stephenson: *The Coastal Plain of North Carolina; the Cretaceous, Lafayette, and Quaternary formations*, N. C. Geol. Econ. Surv., vol. 3 (1912) p. 169-171, 194-196.

W. C. Mansfield: *An oil-prospecting well near Havelock, North Carolina*, N. C. Dept. Conserv. Devel., Econ. Paper no. 58 (1927) 19 pages.

Carolina borings a slope of only about 10 feet per mile. The geophysical results in Virginia are in agreement, showing a slope of about 30 feet per mile from the Fall Line in the vicinity of Petersburg, to a point about 20 miles east of Cape Henry. Beyond this point, the slope increases somewhat. The crystalline floor declines more rapidly than the increase in the depth of water, indicating that the sediments overlying the crystalline rocks thicken eastward.

(2) A second reasonable interpretation of the geophysical data is that the two zones of probable sediments overlying the basement crystallines represent distinct geologic periods. The upper zone, with seismic wave velocities of 5000 to 6000 feet per second, is less dense and apparently is composed of less consolidated and, therefore, probably younger, sediments than the great thickness of underlying strata, which give seismic velocity values of 6500 to 8500 feet per second.

The lower zone is mainly confined to the submerged portion of the Coastal Plain, although it extends under the land a short distance. Ewing finds it in the lower part of the sections at Fort Monroe and Mathews. It seems that this lower sedimentary zone is not continued far up the slope and nowhere crops out. From its position, it is older than the lowest Cretaceous thus far recognized in the Coastal Plain. What age is to be assigned to the strata of these two zones is necessarily problematical at this time. The studies thus far made in the Atlantic Coastal Plain indicate a long break in the sedimentary record after the Appalachian Revolution, as Lower Cretaceous sediments rest on pre-Cambrian crystalline rocks where exposed in the vicinity of the Fall Line. Perhaps the great thickness of fairly dense sediments discovered by the geophysical investigations represents sediments of the Triassic and Jurassic periods that accumulated on a portion of the depressed continent of Appalachia. An examination of the log of the Mathews well reveals nothing that would either prove or disprove the pre-Cretaceous age of the lower strata.

If it were found that the rocks of the Newark series have seismic wave velocities intermediate between those of the Coastal Plain deposits and those of the basement crystalline rocks, one might feel that this was supporting evidence for the view that a great thickness of strata, in part contemporaneous with the Newark, is present under the Coastal Plain but is everywhere concealed. Of course, it is not believed that it will ever be safe definitely to correlate strata on the sole basis of seismic wave velocities, as the densities of sediments are dependent upon many factors other than age. It is of interest to realize further that perhaps marine Triassic and Jurassic sediments accumulated along the Atlantic border in a somewhat similar manner to those of the Pacific Coast. For these reasons, the lower, and larger, portions of the sediments overlying the crystalline rock

floor of the Continental Shelf are provisionally assigned to the Triassic and Jurassic, and it is suggested that they are of marine origin.

The upper zone of the submerged Coastal Plain can readily be assigned to the Cretaceous and Cenozoic, as the wave velocities correspond to those obtained along the line between Petersburg and Fort Monroe where these rocks crop out. Study of the strata outcropping in the Atlantic Coastal Plain, as well as those penetrated by deep borings, has seemed to reveal no sediments on top of the crystalline rocks of pre-Cretaceous age. The lower portions of some of the deep borings may be pre-Cretaceous, although, up to the present, no evidence of such age has been proved, and it is generally believed that the emerged Coastal Plain is composed entirely of Cretaceous and younger sediments. On the basis of the vertebrate fossils, part of the strata now referred to the Lower Cretaceous were at one time doubtfully classified as Jurassic.

(3) A third geological interpretation is offered to account for the geophysical conditions at Camp Lee and at Youngbloods Store. The vibration velocities at those two places were 12,000 to 13,000 feet per second. These seem too high for the Cretaceous and Tertiary deposits and too low for the basement complex of crystalline rocks. It is suggested that, at these places, there may be an area of Triassic (Newark) rocks now concealed by the Coastal Plain formations. Darton [6] records an instance of this sort, where Newark rocks were encountered in a well at Florence, South Carolina, beneath 1335 feet of Coastal Plain sediments. So far as known, no deep well has penetrated to this area of hypothetical Triassic rocks in Virginia.

(4) In addition to the foregoing interpretations, these new geophysical observations supply data that can be used in the consideration of some of the more theoretical problems of North American geology. Among these is an explanation of the manner of disappearance of Appalachia. This theoretical land mass, which paralleled the Appalachian geosyncline and contributed 30,000 to 40,000 feet of sediments that now form the rocks of the Appalachian Mountains, is thought to have foundered shortly after the close of the Paleozoic era. Barrell [7] concluded that the watershed of Appalachia lay about where the 100-fathom line is now, 80 to 100 miles from the present shore line. If the watershed was near the center of the land mass, Appalachia, according to Barrell, was about 200 miles in width. Schuchert [8] is inclined to think that "Appalachia extended eastward even farther than 250 miles" from the present shore line. Accepting these views,

[6] N. H. Darton: *Artesian well prospects in the Atlantic Coastal Plain region*, U. S. Geol. Surv., Bull. 138 (1896) p. 218.

[7] Joseph Barrell: *Upper Devonian delta of the Appalachian geosyncline*, Am. Jour. Sci., 4th ser., vol. 37 (1914) p. 248-249.

[8] Charles Schuchert: *Sites and nature of the North American geosynclines*, Geol. Soc. Am., Bull., vol. 34 (1923) p. 151-230.

which are in general agreement and appear reasonable, the major part of Appalachia is now 10,000 to 12,000 feet below sea level. This conclusion is reached as the result of these new geophysical data showing the present depths of the basal dense rocks, which are most likely the remnants of ancient Appalachia.

It has been a question as to whether Appalachia disappeared by downward warping or by profound faulting. The geophysical data, although admittedly meager, suggest that it sank through warping, inasmuch as the slope of the basement crystalline floor, supposedly once the surface rocks of Appalachia, seems to be fairly uniform, although steepening eastward.

If it is correct to assign the bulk of the sediments of the submerged Coastal Plain to the Triassic and the Jurassic, then the western part of Appalachia remained as land during these periods and contributed sediments eastward. These, together with Cretaceous and Tertiary deposits, have built up the Continental Shelf.

Whether to account for the steeper slope at the margin of the Continental Shelf by faulting, or to compare it with the outer margin of a delta, cannot be satisfactorily determined with the existing information.

Now that there is a method by which ancient rocks can be distinguished from younger ones, it is important that similar geophysical investigations be continued over the deep ocean bottoms. This appears feasible, although it will be a difficult and expensive problem. Perhaps something concerning the Paleozoic deposits that must have formed on the eastern shores of Appalachia may also be learned. Due to prevailing westerly winds that swept across the ancient continent, the rainfall was probably heavier on the western slopes than on the eastern sides, as is usually the case where the prevailing westerly winds from oceans strike land masses. In this case, the burden of sediments carried into the Appalachian geosyncline should have been considerably greater than that of the eastern slopes, which lie now in the depths of the Atlantic Ocean. Nevertheless, it is believed that, somewhere in the Atlantic Ocean, there are extensive Paleozoic strata formed of debris derived from Appalachia and carried eastward by streams. Many fundamental questions of this sort suggest themselves. Perhaps additional geophysical observations across the entire ocean may enable geologists to answer many of them.

(5) The results of the geophysical observations also throw some light on the question of the submerged canyons of the Continental Shelf along the eastern side of North America, which have been the subject of many investigations and discussions in recent years by Spencer, Hull, Upham, Shepard, Stetson, Daly, Hess, MacClintock, and others. From the work of Ewing and his associates, it seems fairly certain that these gorges of the Atlantic Coastal Plain have been cut entirely in rocks of Mesozoic or

Cenozoic age, which are, in the main, of low density and probably unconsolidated. Several explanations for these canyons have been offered, none of which has won wide support.

The reason for discussing the origin of these submarine canyons is the belief that none of the explanations proposed is adequate and the further belief that the true cause will be determined only when geologists learn more about the rocks and the rock structures of the ocean floor. The submerged canyons seem to have been cut by subaerial streams, but it has been found difficult to explain the great depth to which they have been cut below present sea level. When the submerged canyons of the world have been mapped, it seems probable that geophysical investigations of the character of those described by Ewing, Crary, and Rutherford will give information on which an explanation more rational than those thus far proposed can be formulated. It may be found that the oceanic basins have been the sites of great epeirogenic and orogenic disturbances, which have greatly changed the relationships of the existing continents and the oceans.

Other geological interpretations or implications of the geophysical observations might be offered but perhaps enough have been suggested to call to the attention of geologists the value of such investigations. A technique has been developed that will give satisfactory results in water up to 100 fathoms in depth. It is believed that modifications of the methods used will enable the geophysicists to obtain equally reliable results in much deeper waters; in fact, it seems possible to carry a line of geophysical investigations entirely across the Atlantic Ocean.

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FIGURE 1. POWELL AND CODY TERRACES ON THE NORTH FLANK OF MCCOLLOUGH PEAKS
(W) Slope wash on Powell Terrace; (P) Powell gravel sheet; (F) secondary fans of slope wash on Cody Terrace; (E) tertiary fans on Cody Terrace; (C) Cody gravel sheet. (Photo by Douglas Johnson.)

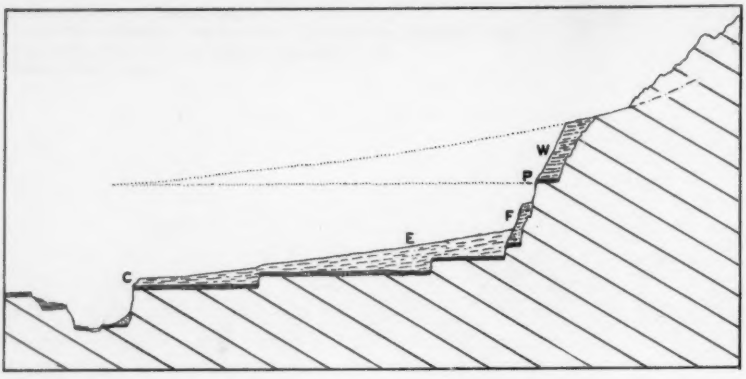

FIGURE 2. SECTION NORMAL TO PLANE OF PHOTOGRAPH

SLOPE WASH ON THE POWELL AND CODY TERRACES

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EROSIONAL HISTORY OF THE BIG HORN BASIN, WYOMING

BY J. HOOVER MACKIN

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INTRODUCTION

GENERAL SETTING

The Big Horn Basin is an elliptical lowland, limited on the west by the Absaroka and Beartooth units of the Rocky Mountain system, and on the south, east, and northeast by the crescentic arch of the Owl Creek, Big Horn, and Pryor Mountains (Fig. 1). The geomorphic history of the basin is divided into two, more or less distinct periods, the first characterized by relative uplift of the surrounding mountain ranges and by widespread fluvial aggradation of the resulting structural depression, and the second by production of the present topographic lowland through partial re-excavation of the Basin fill. The field work on which this paper is based was confined almost entirely to the broad, gravel-capped terraces and interstream benches formed during the second period by degrading streams, and the greater part of the paper is devoted to an analysis of the manner of development and the geological significance of these forms. Major geomorphic problems of the earlier period of mountain-building and aggradation, which bear directly on later degradational history, are treated in a brief preliminary section.

ACKNOWLEDGMENTS

This is a report on one of a series of investigations sponsored by the Big Horn-Beartooth-Yellowstone Research Project,[1] supported in part by a grant from the Penrose Bequest of the Geological Society of America. The field work and preparation of the report were prosecuted under the supervision of Professor Douglas Johnson, who in 1932 as-

[1] W. H. Bucher, W. T. Thom, Jr., and R. T. Chamberlin: *Geologic problems of the Beartooth-Big Horn region*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 167-188.

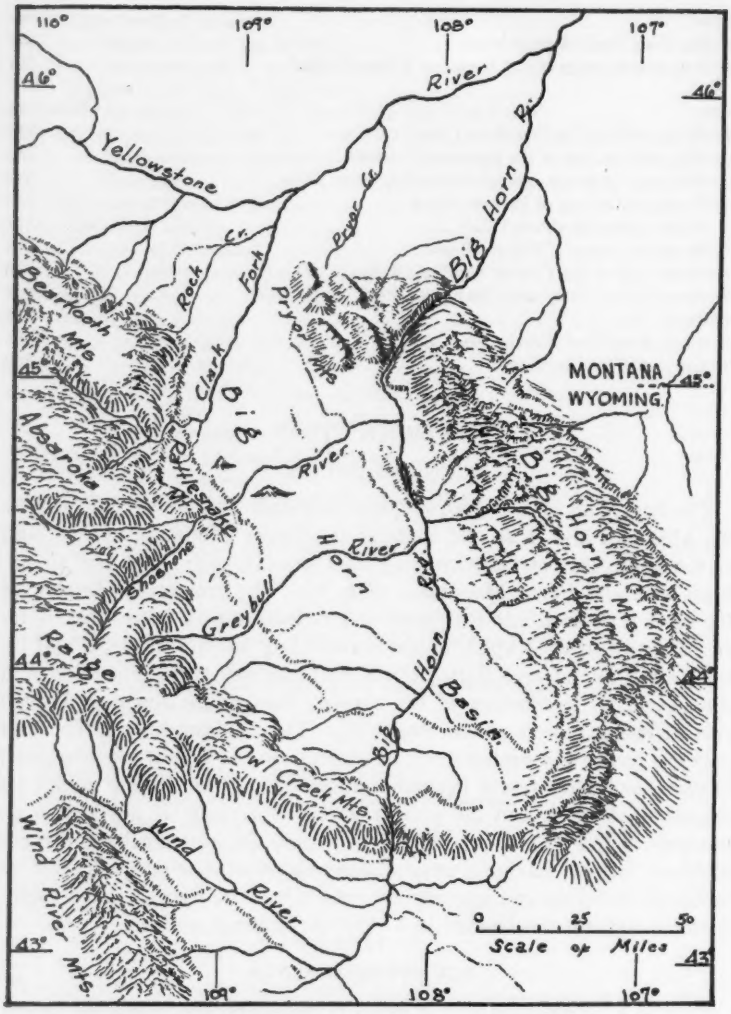

FIGURE 1.—*Physiographic diagram of the Big Horn Basin area*
Drawn by Guy Harold Smith.

sumed general direction of geomorphic studies involved in the Project. It is a pleasure to acknowledge the cooperation of Dr. John Lucke in the mapping of Polecat Bench. Numerous citations in the text fail adequately to express my indebtedness to many other members of the

Big Horn-Beartooth Project; cordial exchange of ideas has been most helpful in every stage of the work. The constant interest and friendly counsel of Professor W. T. Thom deserve special mention.

The several sections of the text have been presented before the seminar in geomorphology at Columbia University and have benefited by the constructive criticism of members of the seminar. The text has been read critically by Professors Johnson, R. T. Chamberlin, R. M. Field, Thom, Dr. W. C. Alden and Dr. H. S. Sharp; but responsibility for interpretations and conclusions rests wholly upon the writer. The physiographic diagram of the Basin was prepared by Professor Guy-Harold Smith, and many of the maps were drafted by Mr. Allen Cary.

Knowledge of the early Tertiary history of the Big Horn Basin is based largely on studies of vertebrate paleontologists and on economic reports of the United States Geological Survey. The problem of the Heart Mountain overthrust and other phases of the structural and stratigraphic history of the ranges surrounding the Basin have been investigated by geologists associated with the Big Horn-Beartooth Project and by other workers. Terraces and benches in the northern half of the Basin have been mapped by Dr. W. C. Alden in the course of his extensive studies of the Montana Plains. The direct bearing of this earlier work on the present study is acknowledged by citations in the text.

PART I: AGGRADATION OF THE BIG HORN BASIN

FORMATION OF THE BASIN BY LARAMIDE DIFFERENTIAL MOVEMENTS

In late Cretaceous time the broad seas that had covered the greater part of western Wyoming and Montana were displaced, and a thick sequence of continental beds was deposited on the Great Plains east of the rising Laramide Rocky Mountains. Deposition gave way to active fluvial erosion over the plains area as a whole during the Eocene, but a practically complete sequence of transitional and early Tertiary sediments was laid down in basins within the mountain front. The fossil record preserved in these sediments in the Big Horn Basin, and the lithology of the detrital materials of which they are composed, taken together, provide evidence bearing on the character of the orogenic movements by which the Basin was formed, and on the general geomorphic aspect of the Middle Rockies of early Tertiary time.

Folding and local unconformities within, and between, the members of the fluvial deposits in the Big Horn Basin suggest that relative uplift of the surrounding ranges, with attendant compression of the basin-filling beds, was more or less continuous during late Cretaceous, Paleo-

cene, and early Eocene time.[2] At present, the difference between the level of the re-exhumed erosion surface at the base of the Paleozoic sequence near the crests of the ranges and the level of this same granitic surface under the basin may be of the order of magnitude of 27,000 feet, the granite floor of the basin being placed at a depth of approximately 19,000 feet below sea level, according to recent geophysical work.[3] The proportion of this structural relief that must be ascribed to pre-Middle Eocene movements is uncertain, for the amount of possible vertical displacements later in the Tertiary is not known; the presence of Paleozoic limestone pebbles in the Fort Union, and granite pebbles in the "Wasatch,"[4] deposits near the borders of the basin[5] proves that the ranges had been greatly raised, by an amount sufficient to permit the removal of the thick Paleozoic and Mesozoic cover from the range cores, at the time these beds were formed.

Continuance of the coal forests of the Cretaceous through the Paleocene indicates that no marked change[6] from late Cretaceous climatic conditions occurred during this interval. The dominant floras of the time were those characteristic of a humid region, with broad swampy areas interspersed by heavily forested lowlands and some open plains.[7] Slow change from the mild climate of the Fort Union coal forests is suggested by red banding of the succeeding "Wasatch" sandstones and shales of the Big Horn Basin,[8] but the overlying Bridger(?) beds[9] contain a flora probably equivalent to that described by Berry[10] from middle Eocene shales from the adjacent Wind River Basin, and likened by him to the present plant assemblage of the South Atlantic and Gulf Coastal States. Tatman Mountain, a high bench in which the lignitic

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[2] W. J. Sinclair and Walter Granger: *Eocene and Oligocene of the Wind River and Big Horn Basins*, Am. Mus. Nat. Hist., Bull., vol. 30, art. 7 (1911) p. 106.

D. F. Hewett and C. T. Lupton: *Anticlines in the southern part of the Big Horn Basin, Wyoming*, U. S. Geol. Surv., Bull. 656 (1917) 192 pages.

W. T. Thom, Jr., and C. E. Dobbin: *Stratigraphy of Cretaceous-Eocene transition beds in eastern Montana and the Dakotas*, Geol. Soc. Am., Bull., vol. 35 (1924) p. 481-505.

G. L. Jepson: *Stratigraphy and paleontology of the Paleocene of northeastern Park County, Wyoming*, Am. Philos. Soc., Pr., vol. 69, no. 7 (1930) p. 463-528.

[3] Beno Gutenberg and J. P. Buwalda: *Seismic methods applied to the Bighorn Basin* (abstract), Geol. Soc. Am., Pr., 1934 (1935) p. 79-80.

[4] For discussion of the present status of the term "Wasatch" see G. G. Simpson: *Glossary and correlation charts of North American Tertiary mammal-bearing formations*, Am. Mus. Nat. Hist., Bull., vol. 67, art. 3 (1933) p. 113-114. The term is used in this paper for the lower Eocene beds in the Big Horn Basin.

[5] D. F. Hewett: *Geology and oil and coal resources of the Oregon Basin, Meeteetse and Grass Creek Basin quadrangles, Wyoming*, U. S. Geol. Surv., Prof. Pap. 145 (1926) p. 30-47.

[6] F. H. Knowlton: *Stratigraphic relations of the Hell Creek beds, Ceratops beds and their equivalents and their reference to the Fort Union formation*, Wash. Acad. Sci., Pr., vol. 11, no. 3 (1909) p. 179-239.

[7] H. F. Osborn: *The Age of Mammals*. New York, N. Y. (1910) p. 100.

[8] F. B. Loomis: *Origin of the Wasatch deposits*, Am. Jour. Sci., 4th ser., vol. 23 (1907) p. 356-364.

[9] W. J. Sinclair and Walter Granger: *op. cit.*, p. 110.

[10] Edward W. Berry: *A flora of Green River age in the Wind River Basin of Wyoming*, U. S. Geol. Surv., Prof. Pap. 165 (1930) p. 58.

beds are preserved, stands now approximately 6000 feet above sea level and 1500 feet above the arid wastes of the central part of the Big Horn Basin. As a warm temperate flora and the associated faunas would not be expected to flourish at an elevation approaching 6000 feet, it appears that the basin floor was at a decidedly lower level during the Eocene than it is at present. Bradley has suggested that the middle Eocene lacustrine sediments of the nearby Green River Basin were deposited not more than 1000 feet above sea level.[11]

These relations support the view that the Laramide orogenic movements in the Middle Rockies were predominantly differential displacements of basins and ranges, absolute uplift of the ranges being accompanied by absolute downwarping of the basin floors. Deposition of the thick Paleocene and Eocene sequences in the basins by streams that were, at the same time, actively eroding on the Great Plains, may have been due, in large part, to these downwarping movements.

The present aridity of Wyoming basins is doubtless related to the altitude of the ranges to the west, which stand in the path of moisture-bearing Pacific winds. The sharp contrast between present climatic conditions and the prevailing humidity indicated by early Tertiary faunas and floras suggests that the ranges of the Middle Rockies did not, in early Tertiary time, reach their present elevations above sea level. Granite detritus in early Tertiary sediments in the Big Horn Basin proves that the encircling ranges had been greatly elevated and deeply eroded, but it does not prove that Laramide differential movements produced high mountains that were peneplaned during the Eocene, for early Tertiary humidity appears to have been a continuation of antecedent Cretaceous conditions rather than a renewal of mild climates after a cycle of uplift and erosion. In harmony with the record of continued mild climates during the period of Laramide differential movement is the suggestion that erosion of relatively weak uparched Mesozoic rocks might have more or less kept pace with uplift of the ranges.[12]

POST-LARAMIDE REGIONAL UPLIFT

The Tatman Mountain lignitic shales are the youngest early Tertiary beds preserved in the Big Horn Basin, but a record of highly significant climatic change is recorded in later deposits in the adjacent Wind River Basin and on the Great Plains. Sinclair and Granger state that the presence of coarse arkoses and limestones of evaporation in the late

[11] W. H. Bradley: *The varves and climate of the Green River epoch*, U. S. Geol. Surv., Prof. Pap. 158 (1929) p. 89.

[12] In a paper that appeared after the first draft of this report was completed, C. Max Bauer bases the same theory on a wholly different type of evidence [*Wind River Basin*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 690-691].

Eocene and Oligocene beds of the Wind River Basin indicates a change from earlier humid conditions of the basin floors in the direction of progressively increasing aridity.[13] Lull and others[14] have shown that the shrinking of forests and the development of true prairies on the Oligocene Plains, as reflected especially in the evolution of the horse, must be ascribed to the same climatic change. By late Miocene time, many of the earlier browsing herbivores of the Plains had become extinct; the types that survived had a dentition adapted for grinding harsh grasses.[15] It thus appears that the semi-arid regime initiated in the Oligocene increased progressively in severity through the mid-Tertiary until, in late Miocene and Pliocene, it approximated the present Plains climate. It is generally agreed, as already pointed out, that the dry climate of the Plains region is due directly to the height of the mountain barrier to the west. For this reason, the slow change from early Tertiary humid conditions, indicated by floras and faunas of the basins, to mid-Tertiary semi-arid conditions on the Plains and in intermontane basins, indicated by widespread changes in all forms of life, is interpreted as the direct result of progressive regional uplift of the Rocky Mountains and the western Plains. There is evidence, noted by many writers, that differential movements of the ranges accompanied regional uplift.[16]

The Eocene streams of the Plains were able, not only to transport eastward the waste that escaped from downsinking intermontane basins, but also to cut deeply into the older Plains strata. If all other Eocene conditions had continued unchanged, regional uplift of the mountains and western Plains in the Oligocene, Miocene, and Pliocene should have caused a continuation or quickening of erosion by eastward-flowing streams through the middle and late Tertiary. Deposition of Oligocene detrital materials in Eocene erosional valleys on the Plains, and the accumulation of middle and late Tertiary deposits over the Central Plains area as a whole indicate that some other factor, which tended to offset this direct effect of uplift, was present. It is believed that this opposed factor was the automatic steepening of the graded declivities of the Plains streams in response to (1) increasing aridity, which involved a loss of water supply, and (2) increasing height of the mountains to the west, which involved a marked increase in amount and coarseness of the load available for transportation eastward. Numerous erosional disconformities in the Great Plains sediments, and the lenselike character and

[13] W. J. Sinclair and Walter Granger: *op. cit.*, p. 114.

[14] R. S. Lull: *The evolution of the horse family, as illustrated in the Yale Collections*, Am. Jour. Sci., 4th ser., vol. 23 (1907) p. 173.

F. B. Loomis: *The evolution of the horse*, Boston (1926) p. 39.

[15] R. S. Lull: *Organic evolution*, New York, N. Y. (1921) p. 614-615.

[16] Eliot Blackwelder: *Post-Cretaceous history of the mountains of central western Wyoming*, Jour. Geol., vol. 23 (1915) p. 112-117.

variable thickness of the stratigraphic units indicate that during the accumulation of the sequence the delicate balance in the graded condition of the streams was repeatedly shifted from deposition to erosion, from time to time, and from place to place. Progressive change from the generally fine-textured channel sands and gravels of the Oligocene to the markedly coarser-textured gravels of the Miocene and Pliocene is independent evidence that the declivities of the Plains streams were progressively steepened during deposition of the sequence.

ALTITUDE OF BIG HORN BASIN FLOOR AT CULMINATION OF PERIOD OF AGGRADATION

Aggradation of the Central Plains area culminated in late-Tertiary time in the production of the vast, eastward-sloping alluvial surface of which the High Plains is a greatly shrunken erosion remnant. Later erosion, especially active near the mountains, has destroyed the former westward extension of the aggraded surface, but the fact that the surface was formed by streams heading in the mountains proves that it formerly extended westward into the intermontane basins, and that these were largely filled with detritus before the initiation of the present cycle of downcutting.

Two lines of evidence, of a direct and local nature, indicate that the Big Horn Basin was filled to a high level, relative to the surrounding ranges, during Tertiary time. The first is the fact that nearly all the major streams cut across structural barriers in positions that indicate superposition from a higher level.[17] The second is the presence of extensive deposits of stream-rounded gravel at elevations from 7000 to 9000 feet on the flanks of the Big Horn Range.[18] The gravel surfaces on the Big Horns slope smoothly outward from the high axial peaks of the range but end abruptly in a steep descending scarp at the range front; projections of the slopes across, and above, the present lowland define in a general way the form of the basin floor at the culmination of the period of aggradation.

The manner of origin of the striking erosion surface at 8,000 to 10,000 feet on the flanks of the ranges that surround the basin is a special problem apart from that under consideration. The surface is generally assigned to the Pliocene and has been called the "Sub-summit peneplain",[19] with the inference that it was formed at a low level (not necessarily at sea level) and that its present elevation and the deep trenching

[17] Nevin M. Fenneman: *Physiography of western United States*, New York, N. Y. (1931) p. 147.
[18] N. H. Darton: *Description of the Cloud Peak-Fort McKinney quadrangles, Wyoming*, U. S. Geol. Surv., Atlas U. S., fol. 142 (1906) p. 8-9, fig. 21.
[19] Arthur Bevan: *Rocky Mountain peneplains northeast of Yellowstone Park*, Jour. Geol., vol. 33 (1925) p. 563-587.

of the present streams are due to uplift subsequent to its formation. But if, as suggested here, the gravel that still mantles part of this erosion surface was formerly continuous with the aggraded floor of the basin and, in turn, with the late Tertiary aggradational surface of the Great Plains, a very different history is suggested. Willard Johnson, in his classic paper on the Great Plains, emphasized particularly a point that has seemingly been overlooked by many later workers in Rocky Mountain geomorphology; namely, that the original slope of deposition of the High Plains Tertiary beds could not have been significantly less than the present eastward slope of these strata.[20] Recognition of this relationship is of critical importance, for it means that the western margin of the Plains, all extensions of the aggraded surface in intermontane basins, and all related erosion surfaces of the mountains could not have been formed near base level, but must have been formed at a high level determined by the declivities of the streams of the Plains and the distance to the sea.

This interpretation, which means essentially that the major regional uplift of the Rocky Mountains occurred during, or before, mid-Tertiary time, is in complete harmony with the record of climatic change on the Plains to the east, for, as already pointed out, the semi-arid to arid regime of the Plains was incipient during late Eocene and Oligocene, became marked in the Miocene, and has continued to the present time. The alternative theory of deep mid-Tertiary degradation of the mountains and very marked late Tertiary or early Pleistocene uplift is in direct conflict with this climatic record. In its relation to the depositional history of the Plains, the theory was satisfactory when the mid-Tertiary sediments were regarded as lake beds, deposited flat and later tilted eastward when the mountains were raised. But Davis[21] and others showed long ago that the Great Plains beds are fluvial, not lacustrine; and Johnson has presented evidence tending to prove that the present eastward slope of the Pliocene beds corresponds closely with their original slope of deposition.[22]

Aside from supposed mid-Tertiary peneplanes of the ranges the most compelling evidence that has been offered in favor of deep mid-Tertiary degradation and later marked uplift is the trenching of the present streams. But Johnson early recognized that the downcutting of the streams in the mountains might be accounted for, in part, by any change, which, by altering their relations of load and discharge, might cause

[20] Willard Johnson: *The High Plains and their utilization*, U. S. Geol. Surv., 22 Ann. Rept., pt. 4 (1902) p. 638.

[21] W. M. Davis: *The freshwater Tertiary formations of the Rocky Mountain region*, Am. Acad. Arts Sci., Pr., vol. 35, no. 17 (1900) p. 372.

[22] Willard Johnson: *op. cit.*, p. 638.

them to regrade their courses across the Plains at a lower declivity.[23]

The quantitative aspects of the problem of change in stream declivity versus regional uplift as an explanation of downcutting are uncertain; no discussion of the matter is attempted here. There is certainly no question but that some regional uplift may have occurred in the late Pliocene or Pleistocene. The writer's purpose is simply to stress the incompatibility of the climatic record of the Plains and the depositional history of the Plains streams with theories of middle or late Tertiary peneplanation of the Rocky Mountains, and to point out that the recognition of a middle or late Tertiary erosion surface at 7000 to 9000 feet in the mountains does not necessarily mean late Tertiary or Pleistocene regional uplift of the same, or nearly the same, amount. These questions bear directly on the reason for downcutting of 3000 to 4000 feet in the Big Horn Basin in the present cycle; they will be discussed further in a section dealing with interpretation of the landforms of the Basin.

BIG HORN BASIN DRAINAGE

The Wind River flows southeastward through the Wind River Basin to a point near its center, where it turns sharply northward and cuts through the Owl Creek Mountains in a deep gorge, to enter the Big Horn Basin as the Big Horn River (Fig. 1). During early Tertiary time the Wind River probably joined the North Platte drainage to flow eastward to the Plains in a consequent course.[24] The diverging Powder, Cheyenne, White, Niobrara, and North Platte rivers of the Plains may be explained as radial consequents inherited from the surface of a broad fan controlled by the Wind River-North Platte system. The peculiar course of the present Wind River is best accounted for by capture by a headward-working tributary of the Big Horn River at some time, before the beginning of the present cycle of erosion, when the Owl Creek Mountains were largely submerged in a sea of waste.

A generalized restoration of the form and drainage pattern of the aggraded floor of the Big Horn Basin may be based on the present stream pattern, which was inherited directly from the surface of the waste fill, and on deductive comparison of the Big Horn Basin with aggraded structural depressions in other regions. The position of the San Joaquin River near the western margin of the Great Valley of California is due to a marked difference in the supply of waste shed into the trough from the high Sierra Nevada mountain block on the east and the much lower Coast Ranges on the west. Similarly, it might be expected that the

[23] Willard Johnson: *The High Plains and their utilization*, U. S. Geol. Surv., 21 Ann. Rept., pt. 4 (1901) p. 655.

[24] W. J. Sinclair and Walter Granger: *Eocene and Oligocene of the Wind River and Big Horn Basins*, Am. Mus. Nat. Hist., Bull., vol. 30, art. 7 (1911) p. 87.

longitudinal stream in the aggraded Big Horn Basin should have been pressed to the eastern margin of the basin by waste shed from the Absaroka Mountains, no higher, but far more extensive, than the Big Horn Range on the east. The present position of the Big Horn River meets this requirement satisfactorily. Its asymmetrical position with respect to the axis of downwarping and its complete disregard of rock structures indicate that the course of the stream in the basin cannot be explained by any theory other than superposition from a position determined by the slopes of a former waste fill. The stream leaves the basin, not by the wide northern gap in the horseshoe of the enclosing ranges, but through a deep canyon cut across the nose of the plunging Big Horn arch (Fig. 1).

The major streams of the Big Horn Basin other than the Big Horn are the Greybull, the Shoshone, and the Clark Fork rivers, all of which issue from the ranges on the west. Near the point where the Shoshone enters the basin, it cuts directly across the plunging nose of the Rattlesnake anticline, in a course similar to that of the Big Horn River on the Big Horn arch. Both streams probably owe their peculiar courses across the mountain arches to superposition, but as differential upward movements may have continued during the period of downcutting, the streams may be, in part, antecedent.[25] The lower courses of all three streams have been changed by capture in the present cycle; the probable relation of the former courses to the slopes of the waste fill will be discussed when the captures are described.

SUMMARY

The lithologic character and fossil content of the late Cretaceous and earliest Tertiary beds of the Big Horn Basin indicate that the floor of the basin stood at a relatively low level during the period of Laramide differential movement by which the structural basin was largely formed. Regional uplift, beginning in late Eocene or Oligocene, raised both the encircling ranges and the prism of inwashed early Tertiary sediments to a high level and caused changes in the character of the east-flowing streams, which resulted in widespread aggradation. Downcutting in the present cycle may be due, in part, to changes in the graded declivities of the streams flowing to the sea. The present drainage pattern of the basin is inherited, with some modification through capture, from the pattern controlled by the slopes of the waste fill.

[25] W. T. Thom, Jr., G. M. Hall, C. H. Wegemann and G. F. Moulton: *Geology of Big Horn County and the Crow Indian Reservation, Montana*, U. S. Geol. Surv., Bull. 856 (1935) p. 84.

PART II: DEGRADATION OF THE BIG HORN BASIN

INTRODUCTION

Discussion of the geomorphic history of an area commonly proceeds from the earliest to the latest stages represented in the region as a whole, this plan of presentation being especially suited to the treatment of peneplanes and other forms of regional extent that are produced by the lowering of interfluves by the combined processes of weathering, wash, and creep. But, as will be demonstrated, the residual landforms of the Big Horn Basin are gravel-capped benches and terraces, produced, not by regional denudation, but by lateral stream corrasion, and each bench and terrace sequence was formed by a single stream, with no necessarily direct relation to the genetically similar sequences of adjacent streams. For this reason, the varied histories of the three major stream systems of the basin (the Shoshone, the Greybull, and the Clark Fork-Rock Creek systems) are presented as distinct units, and the correlation and interpretation of related forms in the several systems are treated in a separate section.

The widespread development of stream-cut, gravel-mantled surfaces in the basin makes a clear understanding of the origin of these forms of paramount importance. For this reason, the usual order of presentation is further altered in discussion of the bench and terrace sequence of the Shoshone River; the forms are described in ascending order from the floodplain, now in process of formation, through higher and successively broader terraces, to the high bench that forms the divide between the present Shoshone Valley and the valley of the Clark Fork River, the adjacent stream on the north.

STREAM-PLANED SURFACES IN PRESENT SHOSHONE VALLEY

INTRODUCTION

In the first 10 miles after leaving its canyon on the west side of the Big Horn Basin the Shoshone River falls approximately 25 feet per mile. This steep gradient is not an average of quiet, graded stretches and local falls or rapids; it represents, rather, the continuous graded slope of a stream nicely adjusted to the transportation of a great load of coarse rock waste. Curved meander scarps, cut deeply into the valley sides at the outside of each sweeping bend, indicate that the vigorous stream is cutting laterally; opposite each cut bank is the floodplain strip that is the direct and necessary result of such lateral cutting. The process of floodplain formation is generally well understood, but the critical importance of floodplain structure in this investigation seems to justify a brief description of the process as it operates in the Shoshone Valley.

ORIGIN OF SHOSHONE FLOODPLAIN

Swift currents, armed with an abundance of rock fragments, impinge on the shale and sandstone banks of the stream at the outside of each curve, and the banks are slowly worn back. The channel cross-section would, thus, be increased but for the fact that, ordinarily, a complementary shoaling and filling takes place on the opposite side of the channel. If the discharge and velocity (and, therefore, the cross-section) of the stream remain constant, deposition can occur no more rapidly than erosion; consequently, during times of normal flow the slack-water side of the channel is slowly filled, the material deposited being the coarsest debris carried or rolled by relatively slow, slack-water currents (Fig. 2, A, f).

During high-water periods, however, the channel section is greatly enlarged; in part, by channel deepening through scouring out of loose bottom fill; in part, by much more rapid cutting on the outside of curves; and, in part, by the scouring away of low-water channel deposits at the inside of curves (Fig. 2, B). As the floodwaters subside and the transporting power of the stream is reduced, deposition again occurs at the inside of bends and in the channel floor, but, in this instance, the rate of deposition from rapidly subsiding floodwaters is rapid, and the debris first deposited is the coarsest handled by the torrent (Fig. 2, A, c). Thus, the gravels of the stream may be effectively reworked and sorted as to size several times before the coarsest of them are finally more or less permanently incorporated in the broadening floodplain.

The floodplain gravel is, therefore, formed as a single sheet by lateral accretion; it is not really a floodwater deposit at all, but a channel deposit; its maximum possible thickness, in a stream that is not aggrading, approaches the maximum depth of effective flood scour. It is often observed, however, that the widening and deepening of the channel during flood periods fails to accommodate the increased discharge, with the result that part, or all, of the gravel sheet is covered by floodwaters. Minor irregularities of the gravel surface and the relative shallowness of the overflow result in a sharp reduction of velocity as soon as the overflow leaves the main current. Fine muds and silts are then deposited from suspension over the whole flooded area, and successive layers of this material are formed in successive high-water stages, until the surface is built up to the highest level of floodwaters (Fig. 2).

Thus, the floodplain is composed of two parts, distinctly different in origin. The gravel represents the coarsest material moved along the stream channel, and is characteristically unstratified. The overlying silts and clays are deposits from suspension, and they represent nearly the finest material carried by the stream. The contrast between the two

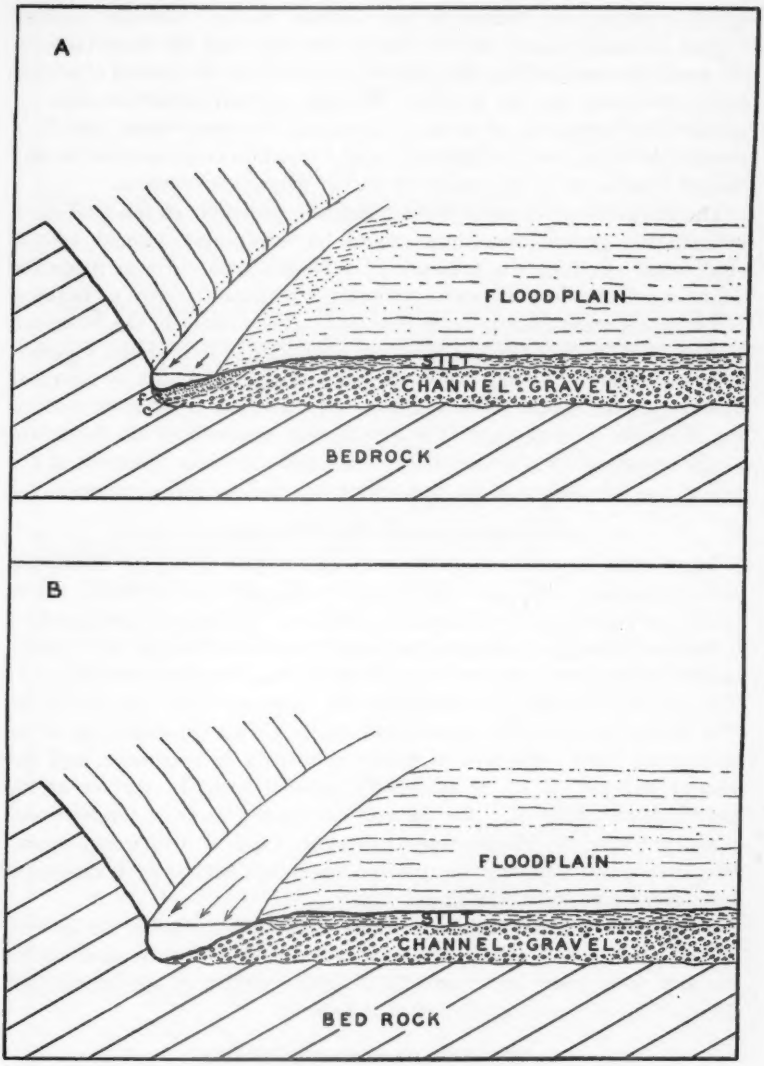


FIGURE 2.—*Formation of the Shoshone valley floor*

(A) Low-water stage: f, fine detritus, laid down in the stream channel during periods of normal flow; c, coarse detritus, laid down in the channel at the end of a period of high water. (B) High-water stage.

layers is particularly marked in the Shoshone Valley, where the channel deposit is mainly black volcanic gravel derived from the mountains to the west, whereas the fine silts, derived in part from the erosion of basin-filling sediments, are tan in color. No such marked distinction may be possible in floodplains of streams handling only fine waste, but it is reasonable to suppose that both types of deposition enter into the formation of floodplains of all stable, or slowly degrading, streams.

The thickness of the gravel sheet and the character of the rock floor beneath the present floodplain cannot be determined without special equipment. In fact, the presence of a beveled rock floor at moderate depth cannot be demonstrated conclusively without test pits or borings, but it seems reasonable to infer that such a floor exists in the Shoshone Valley, from the fact that the Shoshone floodplain is formed and widened, as already shown, through the cutting back of the valley side by a stream of some limited depth. If such a floor does exist and if the river were to cut below the level at which it is now flowing, remnants of the floodplain might remain as steps on the valley sides, and the whole thickness of the gravel and the surface of the underlying bed rock would be exposed.

ORIGIN OF SHOSHONE TERRACES

Many such step-like remnants are exposed just above the floodplain in the Shoshone Valley, and their frontal scarps and the transverse valleys of tributary streams provide excellent sections. The gravel sheet averages 8 feet in thickness, is unstratified, and is composed largely of volcanic pebbles that could not have been derived from the local country rock. The gravel is overlain by floodplain silt, approximately one foot deep. The surface on which the gravel rests neatly bevels the structures of the underlying rock, regardless of minor differences in resistance, and the planed bedrock surface is essentially parallel with the surface of the gravel sheet. Evidently, the terraces are simply parts of former floodplains produced by the companion processes of lateral cutting and deposition, which are widening the present floodplain, and later transformed into terraces by downcutting.

Description of the Shoshone terraces must be preceded by an account of one other process, which, although not directly related to main stream activity, is of great importance in the conformation of the terrace surfaces. This is the accumulation of slope wash.

SLOPE WASH ON SHOSHONE TERRACES

When the laterally planing main stream swings against one of its valley sides, a scarp is produced, and the lower courses of all the tributary streams of the valley side are there cut away (Fig. 3, A). As soon as the main stream retreats from the undercut bank, talus and other locally

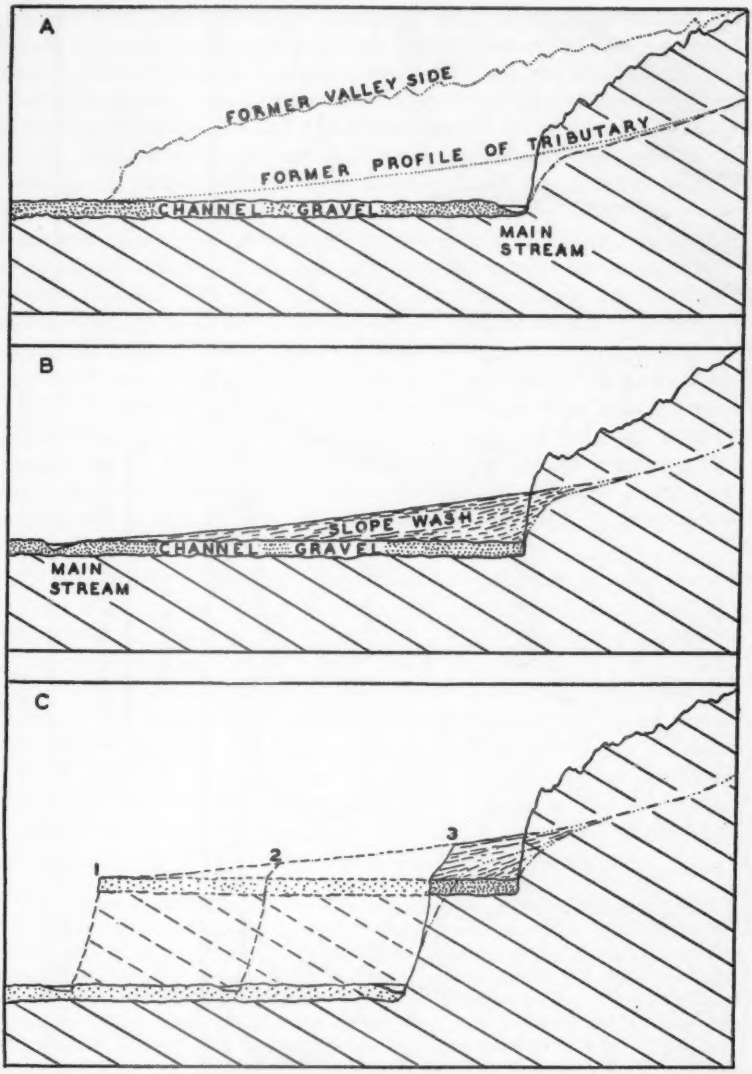


FIGURE 3.—*Origin of slope wash on the Shoshone Terraces*

(A) Truncation of tributary by lateral cutting of main stream; (B) accumulation of slope wash over floodplain gravel sheet after retreat of main stream from undercut bank; (C) effect of slope wash on elevation of terrace surface after incision of main stream.

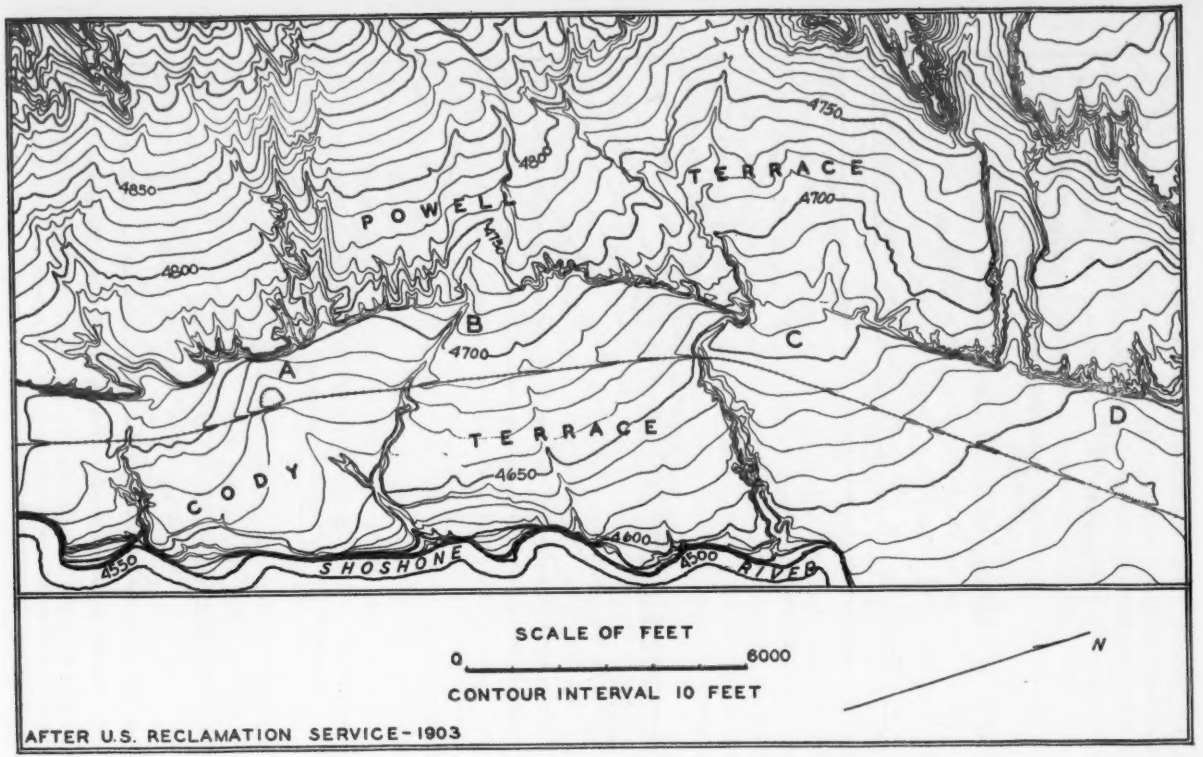


FIGURE 4.—*Slope wash on the Shoshone Terraces*

Map showing a part of the north side of the Shoshone Valley, 10 miles west of Powell. The scarp between the Powell and the Cody terraces is completely masked by alluvial fans of side streams at B and C.

derived waste begin to accumulate at its base, tending to round out the sharp angle cut by the main stream. The truncated tributaries extend their courses across the widening surface of planation, but, because this surface is almost invariably less steep than the profiles of the tributaries, their velocity is checked, and deposition occurs at the mouth of each minor valley. Aggradation continues until a slope, adjusted to the transport of average tributary load by average tributary discharge, is extended from the foot of the scarp to some local base level, as the main stream, the resistant edge of the gravel sheet, or some other tributary (Fig. 3, B). Aggradation ceases and the fans are dissected as soon as the local base level is lowered. The term "slope wash", or simply "wash", is used hereafter to designate all the materials spread over the terrace or floodplain gravel from higher valley sides, including the detritus laid down by concentrated runoff from definite ravines and that washed from the scarp face in inter-ravine stretches. The term may conveniently be reversed to designate the resulting "wash slopes", which are the most striking minor features associated with Big Horn Basin stream terraces. The wash slopes are essentially graded slopes of transportation, formed by aggradation of side streams after an interruption of their normal development by main-stream undercutting.

The degree to which wash may alter the conformation of a terrace surface is well shown by the accompanying map of a portion of the Cody Terrace in the Shoshone Valley (Fig. 4). The terrace actually consists of two, or more, broad, river-cut rock steps, each essentially flat in cross-valley profile, and each mantled by Shoshone gravel; but spreading fans of wash have so completely covered the main rock steps and gravel veneer that these can be seen only in the valleys of tributary streams. The minimum vertical distance between the highest gravel sheet of the Cody Terrace and the tread of the next higher Powell Terrace, as shown by the height of the scarp at points A and D (Fig. 4), is at least 90 feet, but at the fan apices B and C the scarp is completely masked.

Still thicker accumulations of wash are shown in the photograph of the Shoshone terraces on the northern flanks of the McCollough Peaks area (Pl. 1). The Shoshone River and a strip of present floodplain are in the foreground. At the top of the scarp, just above the trees, a sheet of black Shoshone gravel (C), approximately 12 feet thick, rests on beveled bed rock (the Cody Terrace). The top of the black band (P) in the distant scarp is the surface of a gravel sheet of similar thickness (the Powell Terrace), but apparently much thicker, because a talus of resistant gravel covers the bed rock, as in the left half of the Cody Terrace in the foreground of the photograph. A great fan of slope wash (W), spread out by side streams across the broad Powell valley floor, has been truncated by the lateral swinging of the Shoshone at a lower level (restoration suggested by dotted lines in the diagram). Secondary fans (F) have been sectioned by repeated lateral migrations of the stream at the Cody stage, and tertiary fans (E) are now forming on the Cody

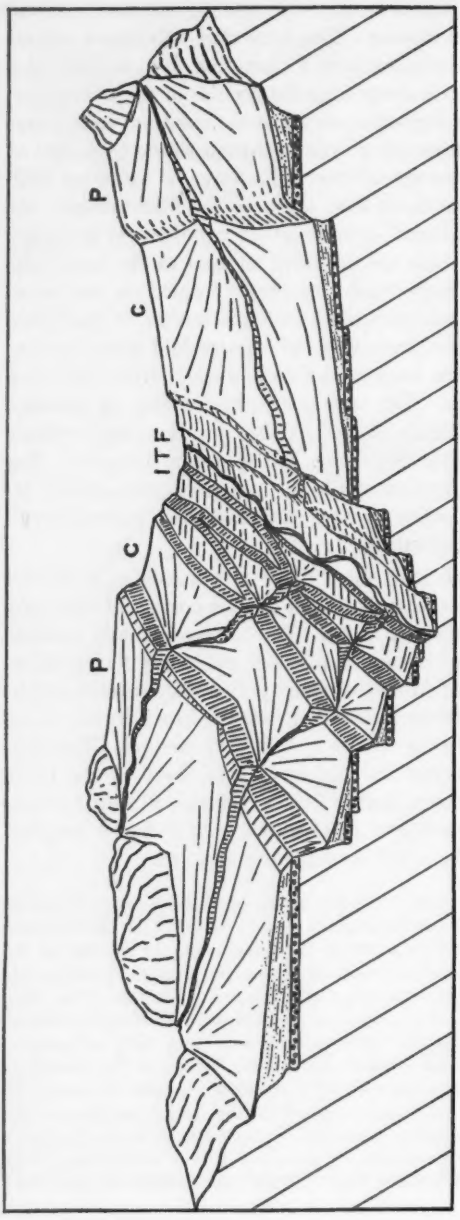


FIGURE 5.—*Shoshone Terraces near Cody*

Generalized diagram, showing stream-cut terraces in the vicinity of Cody, Wyoming. (P) Powell Terrace, (C) Cody Terrace, (ITF) Inner Terrace Flight.

Terrace. The thickness of wash over the main stream gravel in the frontal scarp of the Cody Terrace is approximately 20 feet, and in the distant Powell Terrace scarp, over 150 feet.

The wash mantle, when originally formed on any floodplain or terrace, thins to nothing at its riverward edge (Fig. 3, B), but subsequent cutting back of the frontal scarp may section the wash slope at any point and thus give any thickness of wash at the front edge of the remaining part of the terrace (Fig. 3, C). It is readily apparent, therefore, that the elevation of the topographic surface of a wash-covered terrace, either at its front edge or at its inner margin, is of little significance as to the level of main stream swinging at which the terrace was formed.

The elevation of the main stream gravel sheet itself is the single critical element in terrace correlation; fortunately, the marked contrast between the coarse black gravel of the Shoshone River and the fine-grained, buff-colored wash makes this critical line easily traceable. The figures given as terrace elevations in the text of this article, or plotted in the construction of longitudinal terrace profiles, refer, without exception, to the elevation of the top of the main stream gravel sheet; these elevations, determined by leveling, may differ by 100 feet, or more, from the elevation of the topographic surface of the terrace (Pl. 1).

CODY CROSS-VALLEY PROFILE

In the vicinity of Cody the Shoshone valley sides immediately above the present stream are interrupted by from three to as many as eight ascending topographic steps, grouped here for simplicity as the "Inner Terrace Flight" (Fig. 5). Each terrace is mantled by a 7- to 12-foot layer of coarse Shoshone gravel, which rests on a smoothly sloping surface that bevels bedrock structures and is overlain, in turn, by floodplain silt and a variable thickness of wash. The treads of the terraces of the Inner Flight are generally narrow, are scroll-shaped in plan view, and are not paired on opposite valley sides.

At a point generally about 160 feet above the river, these narrow lower terraces give way to a much more extensive surface, on which the city of Cody is built. The Cody Terrace is really a composite feature, for sections in transverse stream valleys prove that its gently riverward-sloping surface is formed by a mantle of wash, which completely conceals two, or more, broad low steps, each consisting of a uniform sheet of Shoshone gravel resting on bed rock.

Approximately 90 feet above the highest gravel sheet of the Cody Terrace are remnants of a still older and broader surface, called the Powell Terrace, from the town of that name built on its surface near the center of the Basin. The Powell Terrace is identical in structure with the Cody Terrace and the narrower terraces of the Inner Flight, consisting of the typical gravel layer resting on a beveled rock floor and overlain by spreading fans of locally derived waste. It differs from all the lower terraces in this important respect, that the terrace gravel sheet, exposed in the valleys of tributary streams and in scores of miles of the frontal terrace scarp, appears to be a single layer that is at essentially the same elevation on both sides of the valley.

Consideration of the Shoshone terraces in the Cody cross-valley profile serves to bring out clearly the origin of these features. The lithology of the terrace gravel indicates that it could not have been derived from the underlying bed rock. The uniformity in thickness of the gravel sheets and the planed character of the surfaces on which they rest prove that the gravel could not have been deposited during many recurrent periods

of aggradation on surfaces of low relief produced by denudation during many distinct cycles, each shorter in duration than the last. The several terraces are identical in structure with the present floodplain, and it is reasonable to believe that they were formed in the same manner—that is, by the lateral corrasion of the Shoshone River. The terraces differ from each other and from the floodplain only in breadth of development, which (other things being equal) is a function of the time that lateral corrasion continues at the same, or nearly the same, level. Thus, the Powell Terrace represents a pause in downcutting of the Shoshone River, of sufficient duration for the opening out of a valley floor, from 3 to 4 miles broad, by lateral cutting at a single level. The "Intermediate" terrace (Pl. 2) and the broad, down-stepping treads of the Cody Terrace were formed during a later period, when vigorous lateral corrasion was accompanied by slow downcutting. The narrow terraces of the Inner Flight indicate, in the aggregate, a large amount of lateral corrasion, the effect of which is largely lost because quickening vertical corrasion prevented wide planation at any one level, and because the different levels are separated by vertical scarps of considerable height. In the production of the terrace sequence the stream has certainly swung laterally very many more times than are represented by individual treads, for, commonly, only the lateral margins of the widest valley floors are preserved from undercutting by the stream as it swings laterally at lower levels.

It is theoretically certain that, if the Shoshone River were to cut laterally at its present level for a sufficiently long time, it could carve out a valley floor comparable in width to the broad Powell stage surface already described. Similarly, if the change in controlling conditions that caused down-cutting from the Powell level had been longer deferred the Powell stage valley itself would have been still further widened. The conclusion seems inescapable that, given time, a number of adjacent streams such as the Shoshone might open their valleys so widely that the last interstream divides would be cut away and a broad surface of planation, or pediment, be produced. The presence of undoubted river gravels resting on beveled surfaces of planation that commonly form high benches on divides in the Big Horn Basin thus becomes of special significance.

DOWNVALLEY EXTENSIONS OF TERRACES OF THE CODY CROSS-VALLEY PROFILE

The Shoshone terraces, from the Rattlesnake Mountain gorge, at the western margin of the Big Horn Basin, were mapped by the writer for approximately 50 miles, to the Shoshone-Big Horn junction (Pls. 1, 2, 5). A soil survey of the Shoshone Valley provided an excellent base map,[26] and a complete set of unpublished plane-table sheets of a detailed topo-

[26] T. J. Dunnewald and others: *Soil survey of the Shoshone area, Wyoming*, U. S. Dept. Agr., Bur. Chem. Soils, ser. 1927, no. 38 (1927) soil map.

graphic survey, prepared by Army engineers for the Shoshone Irrigation Project, covering the north valley side in the western half of the basin, was of invaluable assistance. The extent of main-stream gravel sheets, rather than the topographic expression of the wash-mantled terrace treads, was the chief concern in mapping; where the inner margins of the gravel sheets were not exposed in tributary valleys, the lateral extent of the former valley floor was inferred from the alignment of truncated spurs rising above the wash slopes and from other topographic relations. Correlation of the terraces was based on walking out the individual levels, and on transverse profiles of the valley, run with plane-table and telescopic alidade, at intervals not greater than 4 miles, the elevation of the several main-stream gravel sheets being noted in each profile. Vertical control was taken from government bench marks directly, from railroad profiles checked against bench marks, and from lines of levels carried along the terrace surfaces between widely spaced bench marks. Essentially the same procedure, with exceptions as noted, was followed in the study of terraces in the valleys of other Big Horn Basin streams.

Small terrace remnants are preserved at several levels more than 300 feet above the Shoshone River, in the eastern half of the basin, but none is sufficiently continuous to permit correlation for any considerable distance along the valley. The highest of the remnants (the Kane Terrace), standing 420 feet above the Shoshone near its junction with the Big Horn River, has a bearing on the date of the diversion of the Shoshone River to its present valley, and will be discussed in a later section.

The Powell Terrace is continuously preserved on one or both sides of the Shoshone Valley in the area under consideration (Pl. 2). The Powell gravel appears to have been one continuous sheet throughout the length of the valley, essentially flat in cross-valley profile and resting everywhere on beveled bed rock. Its thickness varies from an average of 18 feet near the western margin of the basin to an average of 14 feet near the Big Horn River; extreme variations from these averages are less than 5 feet in observed exposures. The Powell stage valley floor decreases markedly in elevation above the present stream as it is traced eastward across the basin, standing 300 feet above the stream at Cody, 120 feet near the center of the basin and 80 feet near the Shoshone-Big Horn junction (Pl. 2, profiles).

As already pointed out, the next lower, or Cody, terrace is really a composite feature, made up of several broad, downstepping gravel sheets formed during a period when vigorous lateral corrasion was accompanied by slow downcutting. The transition to the narrower, higher steps of the Inner Terrace Flight is generally gradual, and the assignment of any local, borderline tread to either group is purely arbitrary. Usually, as

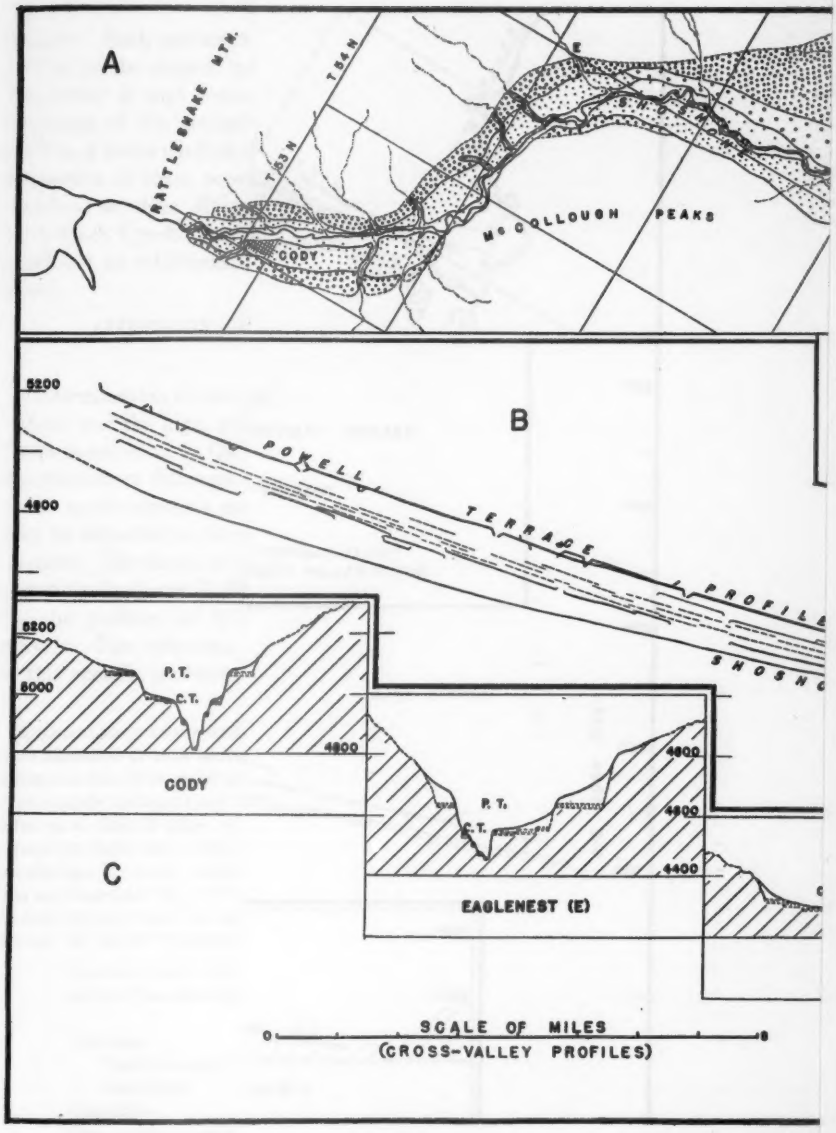
a matter of convenience in mapping, the line was drawn between the broad low steps that are mantled by a single smooth sheet of slope wash and the narrower treads whose higher scarps find surface expression (Fig. 5).

Because each of the Cody Terrace gravel sheets is more or less local, precise correlation of the remnants of any isochronous river-cut surface for any considerable distance is impossible. Moreover, as different gravel sheets crop out from place to place in the frontal scarp, the elevation of the Cody Terrace as a whole cannot be stated by a single figure, as is possible in the case of the Powell Terrace. In spite of this difficulty, the broad Cody valley floor, with the river entrenched now on one side, now on the other, is definitely recognizable and can be traced from the western margin of the Basin to the Shoshone-Big Horn junction.

At the western margin of the basin the vertical range between the several Cody gravel sheets is approximately 90 feet, the lowest being 120 feet above the stream and the highest 210 feet above the stream, or 90 feet below the Powell Terrace. Near the center of the basin the highest gravel sheet is still approximately 90 feet below the Powell level, but the vertical range between the individual Cody gravel sheets has decreased, and the lowest sheet is only 22 feet above the stream. The Cody Terrace is continuous at this elevation to the Big Horn-Shoshone junction.

The Inner Terrace Flight is represented by a nearly straight-walled gorge at the western margin of the basin, where the Shoshone issues from its canyon; there is no floodplain. Three miles farther east, at Cody, the Terrace Flight is well developed, and there is a narrow floodplain; but within the next 30 miles the Terrace Flight disappears, in part by merging with the broadening floodplain, and in part, perhaps, with the not everywhere clearly distinguished Cody Terrace. Lovell is built on a broad Cody stage valley floor whose flatness indicates that the vertical range between the individual gravel sheets near the city of Cody has completely disappeared. This suggests strongly that what was a period of slow downcutting near Cody was a period of relative standstill near Lovell. The marked difference in the post-Cody downcutting of the Shoshone in the eastern and the western parts of the basin is indicated by the fact that the city of Cody is nearly 200 feet above the present river level, whereas Lovell, on the same terrace, is approximately 25 feet above the stream.

The terrace map and the accompanying longitudinal and cross-valley profiles bring out these relations clearly (Pl. 2). Briefly, the divergence of past and present Shoshone profiles, as they are traced westward across the basin, may mean post-Cody differential up-warping of the western half of the basin, while the eastern half has remained relatively sta-



MAP AND PROFI

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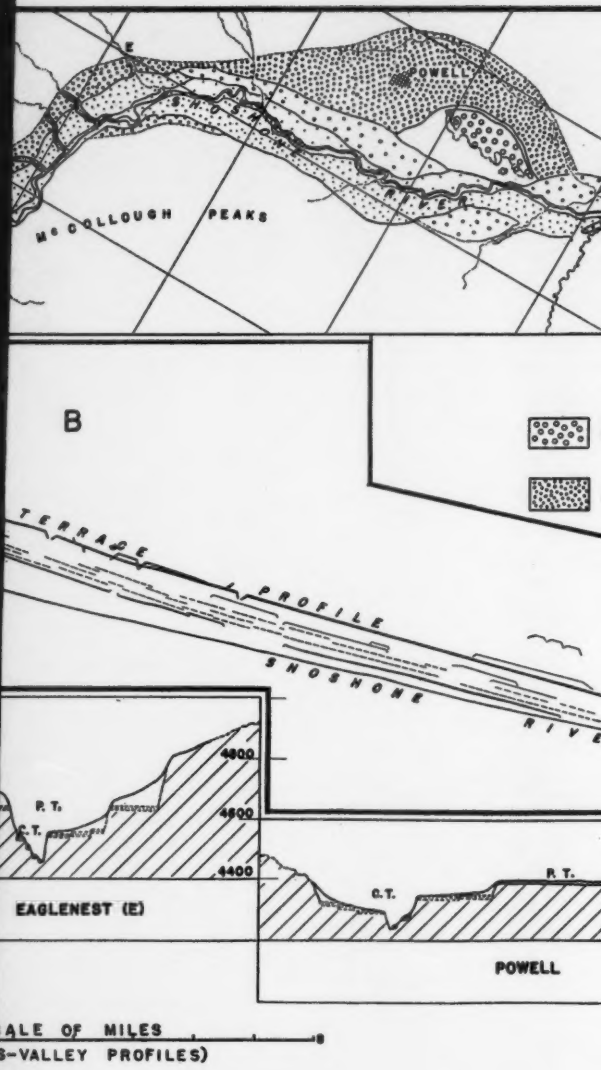

MAP AND PROFILES OF THE SHOSHONE

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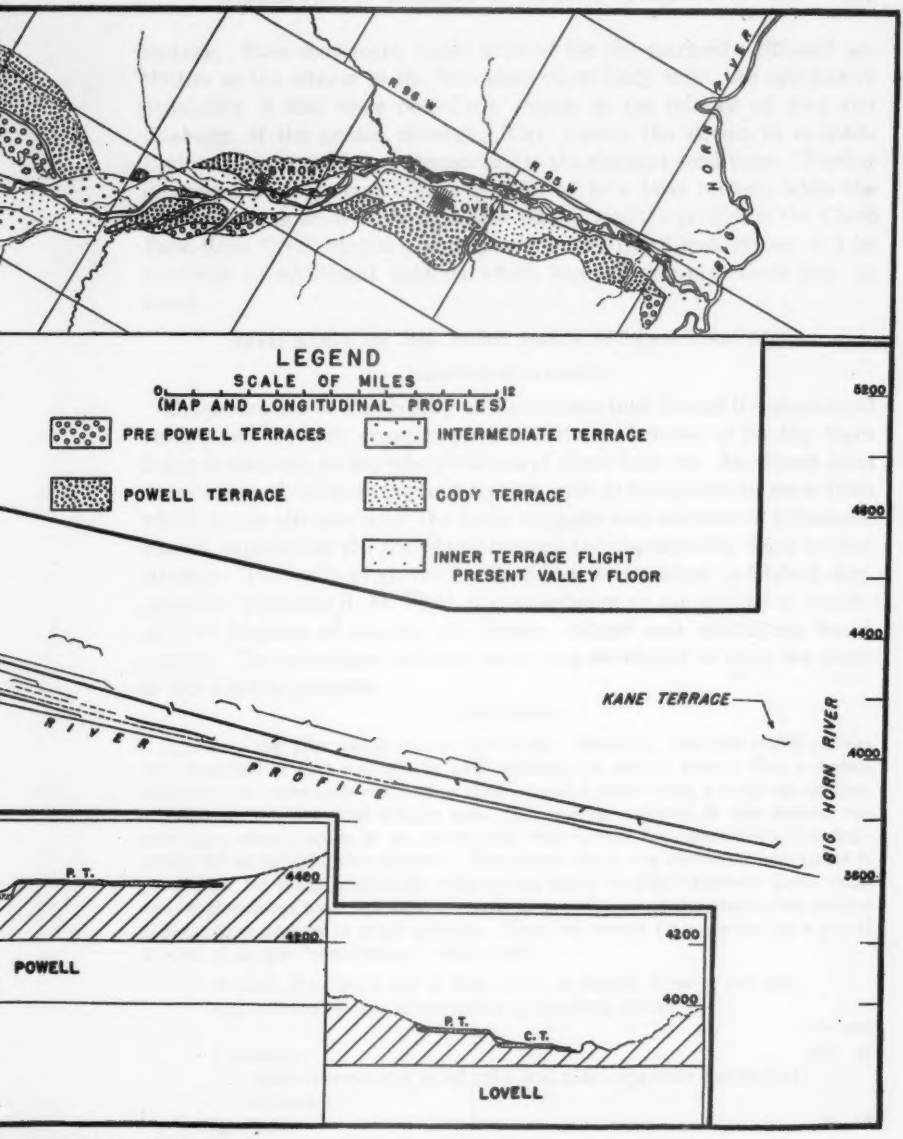


E SHOSHONE TERRACES

les of the terraces, constructed by projecting the levels of onventional symbol, with downward pointing "wings," is used e south valley side point upward. The Powell Terrace profile ants of the Cody Terrace are indicated by solid lines; dashed Cody remnants. The numerous, discontinuous treads of the Shoshone Valley at Cody, Eaglenest (E on map), Powell, and west.

tionary. Such movement would account for the markedly different activities of the stream in the two areas since Cody time. An alternative possibility is that some post-Cody change in the relation of load and discharge of the graded Shoshone River caused the stream to re-grade itself to a lower declivity, appropriate to the changed conditions. Further discussion of these possibilities is deferred to a later section, when the profile of an older, different Shoshone Valley, and the profiles of the Clark Fork-Rock Creek system and of the Greybull-Dry Creek system will be available as additional data on which regional interpretations may be based.

LITHOLOGY OF BIG HORN BASIN STREAM GRAVELS

GENERAL STATEMENT

Determination of the identity of the streams that formed the abandoned valleys and the high, gravel-capped interstream benches of the Big Horn Basin is essential to the interpretation of these features. An almost ideal distribution of different types of country rock in the mountain areas from which major streams enter the basin suggests that distinctive differences may be expected in the gravel transported and deposited by these several streams. The study of gravel "composition" summarized in Plate 3, suggested by Professor R. M. Field, was undertaken as one method of attacking the problem of tracing old stream courses and identifying bench gravels. The technique outlined below was developed to meet the needs of this specific problem.

TECHNIQUE

Approximately fifty pebble counts were made. Generally, from 100 to 300 pebbles were classified at each station. In each instance, all pebbles greater than a certain minimum size were gathered from a mass of gravel taken from a fresh cut or from a previously circumscribed surface area. The sample collected in this manner was piled in a cleared space or on canvas and was divided into appropriate lithologic groups by rapid sight identification. The actual count was converted into terms of percentage, for ready comparison with counts made at other stations. Each count was supplemented by a five-minute search in the vicinity of the station for pebbles of rock types present in small amount. Thus, the record for a station on a gravel deposit of simple "composition" might read:

(Sample from fresh cut in front scarp of Powell Terrace, one mile east of Powell bridge, on north side of Shoshone River)

		Per cent
Volcanics (basic breccias and porphyritic and non-porphyritic basalts and andesites)	188	88
Quartzites	25	12

Five-minutes search of gravel bank:—

3 pebbles gray limestone
1 pebble coarse foliated pink granite
1 fragment petrified wood

EXPLANATION OF SYMBOLS

The many varieties of rock that crop out in the mountains that encircle the Big Horn Basin are reduced on the map (Pl. 3) to four major types best suited to bring out clearly significant differences in the kind of debris supplied to each of the major streams. Each of the bar diagrams except No. 7 represents the average composition of all the fluvial gravel deposits associated with the adjacent stream. The type of rock represented by each of the bars is indicated by a symbol corresponding to that by which the surface outcrop of the same type of rock is defined. Thus, the Shell Creek gravel (No. 11) is made up of approximately 75 per cent sedimentary pebbles and 25 per cent plutonic igneous pebbles. Of the two notably different types of gravel indicated in the Pryor Creek valley, No. 3 represents Polecat Shoshone gravel, predominantly volcanic (compare with No. 8), whereas No. 2 represents the predominantly limestone gravel deposited by Pryor Creek after the diversion of the Polecat Shoshone. Diagram No. 6 represents normal Clark Fork gravel; No. 7 represents glacial outwash from the Clark Fork canyon.

SUPPLEMENTARY STATEMENTS

Sizes of pebbles classified.—In the usual stream deposit, there is every gradation from relatively coarse gravel to fine sand. Separate counts of large and small pebbles at the same station indicate that these different size groups may differ in the proportions of the several lithologies represented. Similarly, counts of lithic fragments and mineral grains in associated sands give proportions notably different from those calculated from rock types represented in the gravel. It is evident, therefore, that the fixing of any minimum size for pebbles classified must introduce inaccuracy, insofar as the true composition of the river deposit is concerned. Composition, in fact, involves the weight, rather than the number, of the fragments of rock types present. Obviously, however, some minimum size must be fixed if pebbles, as such, are to be counted; the convenient minimum of one inch in greatest diameter was adopted in this study. The term "composition", as used here, means the relative number of pebbles (more than one inch in diameter) of different lithologies present in the stream deposit.

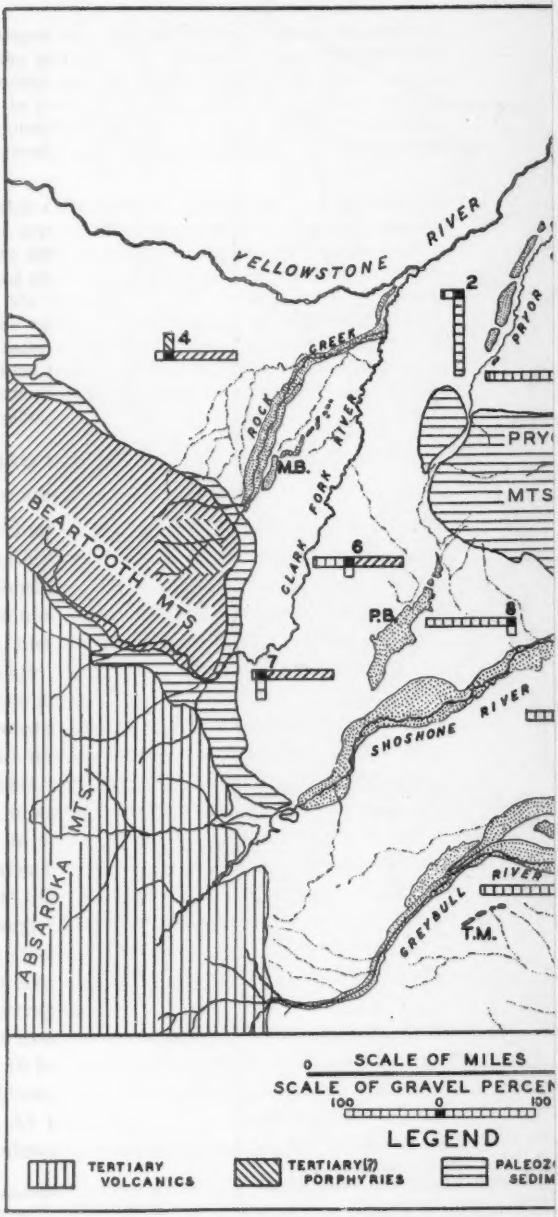
Method of sampling.—Repeated test counts of many pebbles from one, or more, fresh cuts in old, high-level gravel deposits and of pebbles gathered from previously circumscribed surface areas at the same stations show markedly different results. Two counts on the Meeteetse Rim gravel sheet in the Greybull region serve to bring out these differences clearly:

A (*sample from fresh cut in gravel sheet*)	Per cent	B (*sample from gentle, gravel-mantled slope near A*)	Per cent
Volcanics	89	Volcanics	46
Quartzites	11	Quartzites	54

The presence of many thoroughly decayed volcanic pebbles in fresh cuts in the Meeteetse Rim gravel sheet indicates that the difference in the counts is due to selective decomposition of the volcanic pebbles in place. The common occurrence, on the surface of other high benches, of rock cores (surrounded by disintegration debris) of what were formerly boulders of coarse-grained granite, associated with pebbles of fine-grained porphyries that still retain their stream polish, indicates that differences in rate of disintegration of different rock types may also play an important

Symbol

1. Yellowstone flood-plain and terrace gravel.
2. Pryor Creek flood-plain and terrace gravel.
3. Polecat Shoshone gravel in Pryor Creek Valley.
4. Mesa Bench and Rock Creek flood-plain and terrace gravel.
5. Big Horn River gravel northeast of the Big Horn Mountains.
6. Clark Fork terrace gravel.
7. Clark Fork flood-plain gravel and glacial outwash.
8. Polecat Bench and Shoshone River gravel.
9. Big Horn River gravel above the Shoshone Junction.
10. Tatman Mountain and Greybull terrace and flood-plain gravel.
11. Shell Creek gravel.



LITHOLOGY OF STREAM GRAVELS IN THE

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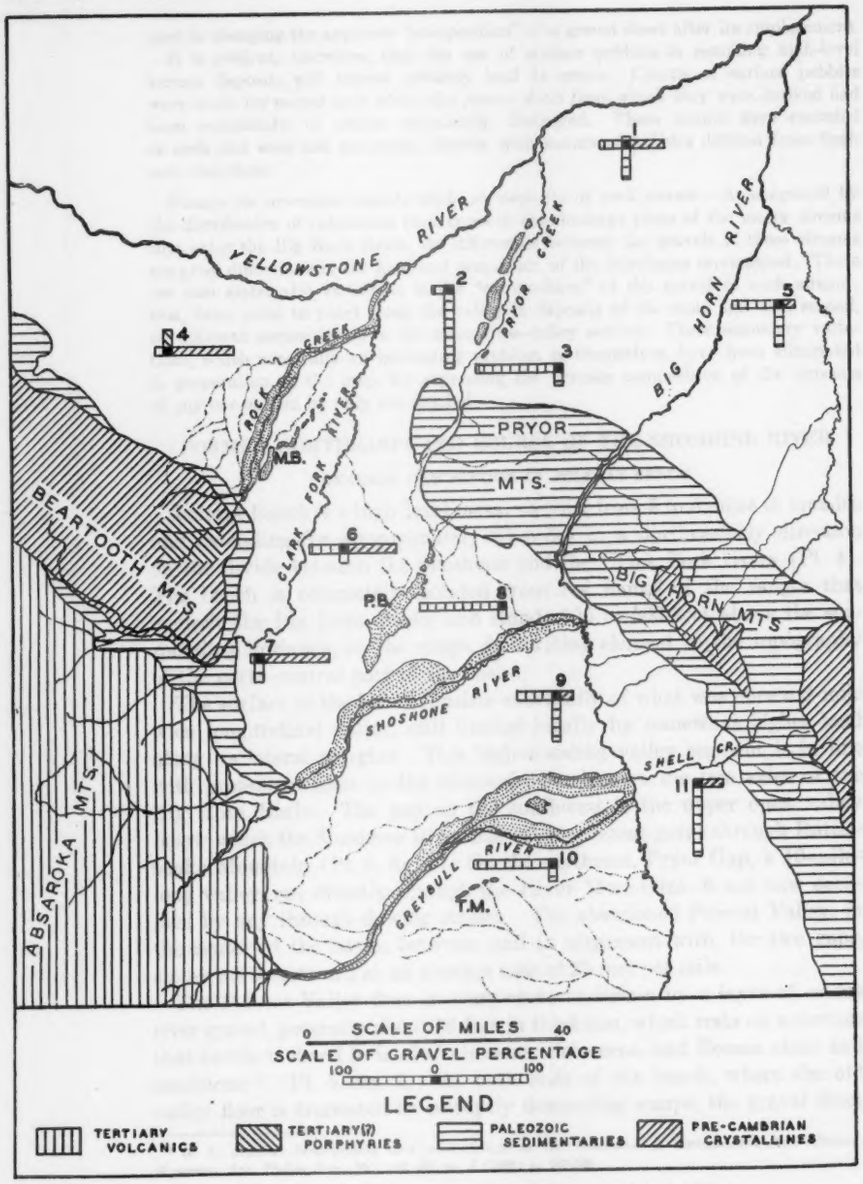


LITHOLOGY OF STREAM GRAVELS IN THE BIG HORN BASIN

Geological boundaries are taken from the United States Geological Survey Geological Map of the United States (1932) except boundaries of Tertiary (?) porphyries in the Beartooth Mountains, which are not precisely known.

part in changing the apparent "composition" of a gravel sheet after its emplacement.

It is evident, therefore, that the use of surface pebbles in sampling high-level terrace deposits will almost certainly lead to errors. Counts of surface pebbles were made for record only where the parent sheet from which they were derived had been completely, or almost completely, destroyed. These counts were recorded as such and were not compared directly with counts of pebbles derived from fresh cuts elsewhere.

Reason for averaging counts made on deposits of each stream.—As suggested by the distribution of contrasted rock types in the drainage areas of the major streams that enter the Big Horn Basin, the differences between the gravels of these streams are gross differences in the kind and proportion of the lithologies represented. There are also systematic variations in the "composition" of the gravel of each stream; first, from point to point along the valley in deposits of the same age, and, second, on different terrace levels in the same cross-valley section. These secondary variations, which constitute an interesting problem in themselves, have been eliminated in preparation of the map, by expressing the average composition of the deposits of any one stream by only one symbol.

FORMER NORTHEASTWARD COURSE OF THE SHOSHONE RIVER

GENERAL DESCRIPTION OF POLECAT BENCH

Polecat Bench is a high-level mesa, varying from 3 to 5 miles in breadth and extending for approximately 20 miles in a northeasterly direction on the divide between the Shoshone and the Clark Fork rivers (Pl. 4). The bench is completely isolated from the flanks of the ranges that encircle the Big Horn Basin, and stands 200 to 500 feet above the surrounding lowlands, as the single dominating element in the topography of the north-central part of the basin.

The surface of the bench consists essentially of what was once a 3-mile wide longitudinal valley, still limited locally by somewhat higher land along its lateral margins. This high-standing valley segment is in line with prominent gaps in the mountain barriers on the two sides of the Big Horn Basin. The gap on the southwest is the upper open valley below which the Shoshone River has cut its present gorge through Rattlesnake Mountain (Pl. 6, fig. 1). On the northeast, Pryor Gap, a 10-mile-long valley, cut directly through the Pryor Mountains, is not now occupied by any through-flowing stream. The abandoned Polecat Valley, in the center of the basin, between, and in alignment with, the two gaps, slopes northeastward at an average rate of 25 feet per mile.

The Polecat Valley floor is everywhere underlain by a layer of coarse river gravel, generally about 16 feet in thickness, which rests on a surface that bevels inclined beds of Cretaceous, Paleocene, and Eocene shale and sandstone[27] (Pl. 5, fig. 2). At both ends of the bench, where the old valley floor is truncated by abruptly descending scarps, the gravel sheet

[27] G. L. Jepson: *Stratigraphy and paleontology of the Paleocene of northeastern Park County, Wyoming*, Am. Philos. Soc., Pr., vol. 69, no. 7 (1930) p. 463-528.

maintains a remarkably smooth surface to the very edge of the break (Pl. 6, fig. 1). Along the lateral margins of the bench, on the other hand, the gravel is generally overlain by a considerable thickness of slope wash. Inward, the wash surfaces slope gently down to the medial flat (Pl. 4, cross profiles; Pl. 6, fig. 2). If the slopes of these surfaces are projected outward, they run up into the air; therefore, they strongly suggest the former presence of higher valley sides (from which the wash was derived) where now steep scarps descend hundreds of feet below the old valley floor. Remnants of this former higher land are still present for considerable distances along the lateral bench margins, and ideal exposures show the gravel sheet, the old inward-facing valley sides, the wash, and the present outward-facing scarp (Pl. 7). As might be expected, terraces marking earlier stages of river-cutting above the level of the final valley floor are developed locally along the valley sides. To complete the picture, areas of higher land, reduced to a mature stage of regional erosion by minor tributaries graded to the old central valley, are still preserved along part of the southeastern and northwestern sides of the bench.

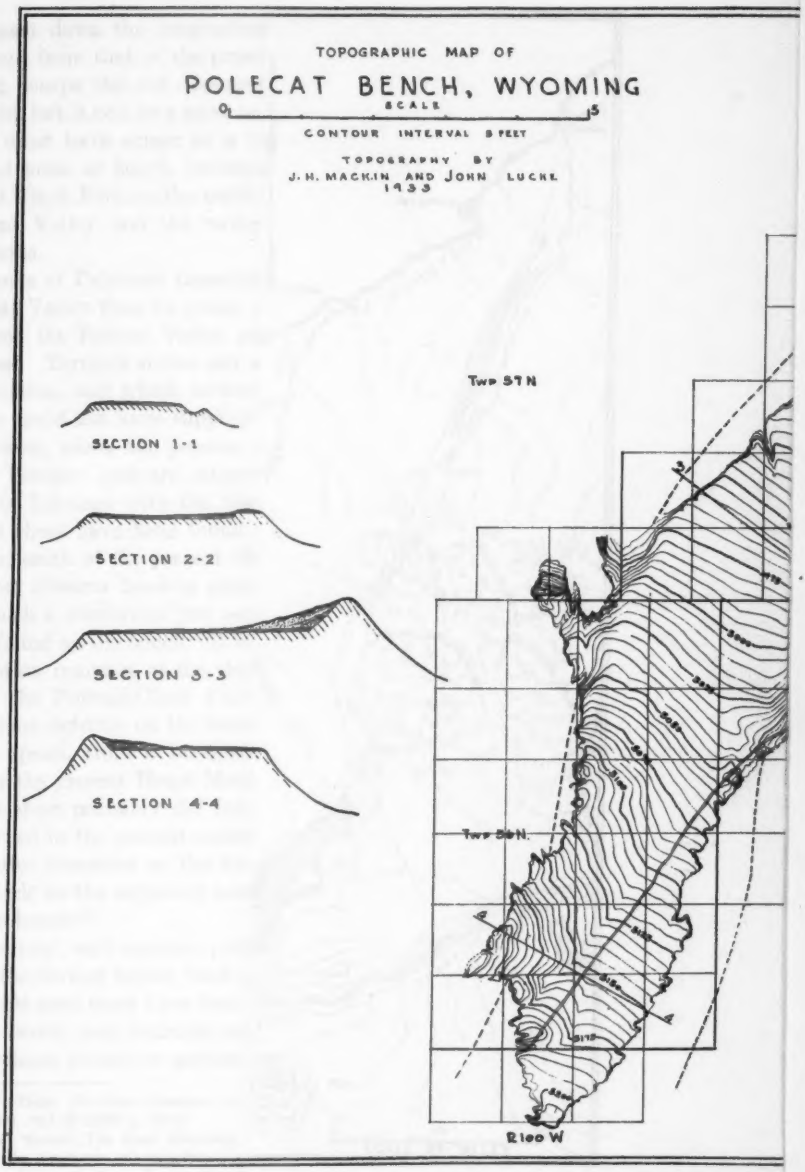
IDENTITY OF BENCH-FORMING STREAM

The position of Polecat Bench on the Shoshone-Clark Fork divide suggests that either one or both of these streams might have formed its surface. Alden, on the basis of the location and the general elevation of the bench, tentatively correlates it with an upper open valley above the present canyon of the Clark Fork.[28] But the Clark Fork, in order to have entered the northeastward-trending valley of the bench, would have had to swing far to the south and then turn sharply back to the northeast. A northeastward-trending segment of the Shoshone River is, on the other hand, directly in line with the axial valley, and the present stream turns sharply to the east, at what is interpreted as an elbow of capture, at a point only 7 miles from the southwestern end of the bench (Pl. 4). Conclusive proof that the Polecat Valley was formed by the Shoshone, not by the Clark Fork, is found in the essential identity in composition of the bench gravels with gravels in the present Shoshone Valley, and in diagnostic differences between these gravels and those of Clark Fork terraces. Plutonic igneous rocks form a large proportion of the Clark Fork gravel, but, during five weeks spent on Polecat Bench in preparing the topographic map, no single plutonic igneous pebble was found (Pl. 3, symbols 6 and 8).

POLECAT STAGE TOPOGRAPHY BEYOND PRESENT MARGINS OF THE BENCH

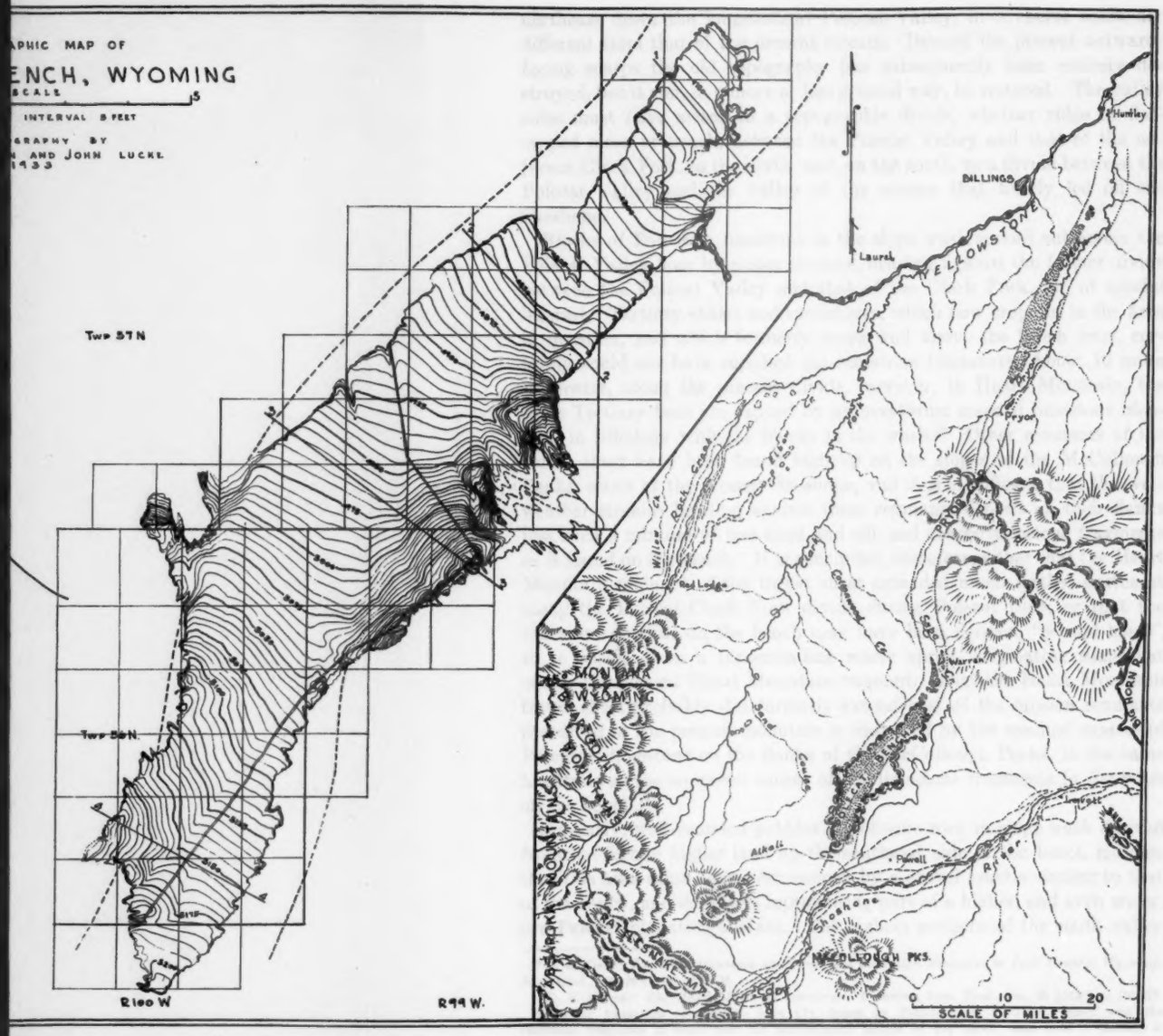
The Polecat Bench surface thus preserves, almost unchanged, a part of the topography of the basin as it existed when the Shoshone flowed

[28] W. C. Alden: *Physiography and glacial geology of eastern Montana and adjacent areas*, U. S. Geol. Surv., Prof. Pap. 174 (1932) p. 26-27.



TOPOGRAPHIC

Dashed lines on topographic map indicate former widt
mate except where they join the margins of the bench,
Hachure map shows the course of the Polecat Shoshone
old valley floor, are approximate in the stretch northeas
Mountain waste. (See Plate 9 and Figure 6.)



TOPOGRAPHIC MAP OF POLECAT BENCH

lines on topographic map indicate former width of gravel-mantled Polecat Shoshone valley floor. Lines are approxi- where they join the margins of the bench, where they are controlled by the type of evidence shown in Plate 7. p shows the course of the Polecat Shoshone River before the Shoshone capture; dotted lines, showing width of the oor, are approximate in the stretch northeast of Polecat Bench, the abandoned valley is deeply filled with Pryor aste. (See Plate 9 and Figure 6.)

northeast down the longitudinal Polecat Valley, in a course markedly different from that of the present stream. Beyond the present outward-facing scarps the old topography has subsequently been entirely destroyed, but it can, in a more or less general way, be restored. The valley sides must have arisen to a topographic divide, whether ridge or flat-topped mesa or bench, between the Polecat Valley and that of the adjacent Clark Fork on the north, and, on the south, to a divide between the Polecat Valley and the valley of the stream that finally led off the Shoshone.

Blocks of Paleozoic limestone in the slope wash spread out across the Polecat Valley floor by minor streams, heading against the former divide between the Polecat Valley and that of the Clark Fork, are of special interest. Tertiary shales and sandstones, which now crop out in the area in question, and which formerly stood well above the bench level, certainly could not have supplied the limestone fragments. About 10 miles southwest, along the present divide, however, in Heart Mountain, the same Tertiary beds are capped by an overthrust mass of limestone identical in lithology with the blocks in the wash.[29] Other remnants of the thrust sheet have been found high up on the slopes of the McCollough Peaks, south of the present Shoshone, and it is significant that the wet-weather streams heading against these remnants deposit on their flanks just such a mixture of fine sand and silt and coarse limestone fragments as is found on the bench. It is concluded, therefore, either that the Heart Mountain remnant of the thrust sheet extended much farther northeast along the Polecat-Clark Fork divide during Polecat time, or that the limestone detritus on the bench may have been derived, "second hand", so to speak, from a far-spreading waste apron, originating somewhat nearer the present Heart Mountain remnant. That the Heart Mountain thrust sheet probably did formerly extend east of the erosion remnants preserved in the present mountain is indicated by the residual masses of Paleozoic limestone on the flanks of the McCollough Peaks, in the same longitude as the supposed source of the limestone fragments in the wash on the bench.[30]

Similarly, well-rounded pebbles of volcanic rock in slope wash derived from the former higher land on the southeast side of the bench indicate that this area must have been capped by a gravel mantle similar to that of the bench, and doubtless representing part of a higher, and even wider, pre-Polecat planation surface. The highest portions of the south valley

[29] C. L. Dake: *The Hart Mountain overthrust and associated structures in Park County, Wyoming*, Jour. Geol., vol. 26 (1918) p. 45-55.

D. F. Hewett: *The Heart Mountain overthrust, Wyoming*, Jour. Geol., vol. 28 (1920) p. 536-557.

[30] Angular fragments of andesitic lava (?), found by Professor O. T. Jones, mixed with the limestone fragments in wash near the northwestern margin of the bench, may have a bearing on the problem of the relations of the Absaroka volcanics to the overthrust sheet.

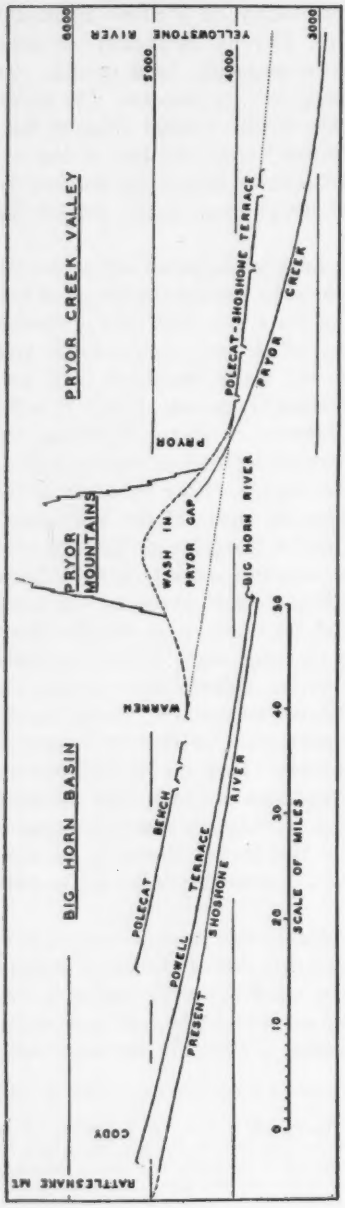


FIGURE 6.—*Profile of the Polecat Shoshone River*

Polecat gravels disappear under Pryor Mountain wash near Warren and re-appear just north of Pryor. Restoration of the profile between these two points is indicated by a smooth dotted line; the higher, dashed, line represents the approximate surface of the wash filling in the abandoned valley. The profiles of the Shoshone River and the Powell Terrace in the present Shoshone Valley have been revolved into the vertical plane of the former Shoshone Valley; the declivity of the Powell Terrace is essentially the same as that of the corresponding segment of the Polecat Shoshone profile.

sides of the bench are, in fact, capped by remnants of this once more-extensive Shoshone gravel sheet.

NORTHEASTWARD EXTENSION OF POLECAT GRAVELS

The gravels of the axial Polecat Valley can be traced northward along the west flanks of the Pryor uplift, toward the Pryor Gap, but before the gap is reached the gravels disappear under a thick mantle of wash derived from the Pryor Mountain slopes (Pl. 4). The present floor of the gap is made up of a series of coalescing alluvial fans, formed by minor side streams, and the bed of the stream that cut the transverse valley is nowhere seen. About 4 miles northeast of the end of the gap, however, in a valley now occupied by the small Pryor Creek, volcanic gravels are exposed. The fact that these basic igneous rock types could not have been derived from the Pryor Mountains, which are composed entirely of sedimentary rocks, mostly limestone, and the essential identity of the gravel, ex-

cept for a slight admixture of limestone, with the gravel traced from Polecat Bench toward the west end of the gap, is apparently conclusive evidence that the Polecat Shoshone did formerly flow through the gap.[31] The volcanic gravels are exposed in broad terraces in the valley now occupied by Pryor Creek, continuously for 25 miles, from the northeast end of the gap to within a few miles of the Yellowstone.

Precise correlation of the Shoshone gravel in the valley of Pryor Creek with the Polecat Bench gravel sheet (in the extension of the longitudinal profile of the Polecat Shoshone River) is made possible by several exposures in the Pryor Creek valley (Pl. 8). These exposures show a lower sheet of gravel, consisting of 80 to 90 percent volcanic rock types and 10 to 20 percent limestone, overlain by another distinct sheet, consisting of approximately 90 percent limestone and 10 percent volcanic pebbles (Pl. 3, symbols 3 and 2). It is believed that the lower gravels were deposited by the Polecat Shoshone River itself, and the upper gravels by the limestone-eroding Pryor Creek, the small percentage of volcanic materials in the limestone gravels having been derived from reworked Shoshone deposits. The dividing line between the two types indicates the moment at which the Shoshone ceased to occupy the Pryor Creek valley and, hence, the exact date of its diversion from the course across Polecat Bench to its present valley. The longitudinal profile of Polecat Bench and other gravel patches west of the Pryor Mountains, and the profile of the major volcanic gravel terrace in the Pryor Creek valley, if connected according to the correlation thus outlined, fall into one smooth curve, the graded profile of the old Shoshone River (Fig. 6).

CAPTURE OF POLECAT SHOSHONE RIVER

Diversion of the Shoshone River from its former course through the Pryor Mountain arch to its present valley might be considered a normal case of the withdrawal of a stream from a hard-rock barrier by a headward-working tributary of a larger and more competent stream (the Big Horn River) crossing a similar barrier (the Big Horn Mountain arch). But if, as already shown, the Shoshone was completely, or almost completely, graded across the Pryor Mountains when the capture occurred, it may be questioned whether the hard rock of the Pryor Mountains did "hold up" the stream sufficiently to cause the capture. It is believed, rather, that the Shoshone was vulnerable to capture by a Big Horn tributary, because its graded declivity, adjusted to the transportation of much coarse rock waste, was (and is) over ten feet per mile higher than that of the Big Horn River. As the capture actually occurred at a point

[31] The theory that the Shoshone River flowed through Pryor Gap at the Polecat stage was advanced, without discussion, in Guidebook 24, *Yellowstone-Beartooth-Big Horn region*, Intern. Geol. Congr., 16th ses. (1933) p. 34.

approximately 40 miles from the Big Horn, the declivity of the small capturing stream must have been sufficiently low (compared with that of the Shoshone) so that the difference in elevation of the main streams was not lost in this distance. It will be demonstrated in a later section that, under the special conditions present in the Big Horn Basin, the declivities of some small basin streams (as the captor) may be as low as, or actually lower than, the graded declivities of much larger streams (as the Shoshone), which enter the basin from the surrounding mountains.

MINOR POST-CAPTURE ADJUSTMENTS ALONG POLECAT SHOSHONE VALLEY

General statement.—The former Shoshone was tapped and led off to the Big Horn at a point nearly 70 miles from its former junction with the Yellowstone. Subsequent changes affecting the abandoned valley, particularly the partition of much of its drainage between the adjacent Clark Fork on the north and the present Shoshone on the south, are of interest apart from the significance of the major capture in the history of the region and may be considered briefly.

Capture of Cottonwood Creek.—Cottonwood Creek rises in a maturely dissected area northeast of the middle portion of Polecat Bench, flows directly southeast for 6 miles, turns abruptly to the northeast to follow a shallow valley cut just below the bench level for about 5 miles, and, finally, makes a second right-angle turn to flow northwestward to the Clark Fork (Pl. 4). It seems reasonable to believe that the creek formerly joined the through-flowing Polecat Shoshone River, that it continued to flow northeastward for some time after the Shoshone capture, and that it was itself, in turn, captured by an aggressive Clark Fork tributary. Other tributaries of the Clark Fork have, in a wholly one-sided contest, led off all the orphaned streams that formerly entered the Polecat Valley from the west and northwest.

Alluviation of Pryor Gap.—A smooth curve, connecting the longitudinal profile of the axial Polecat Valley and that of the correlative terrace in the Pryor Creek valley, should, as shown, approximate the profile of the Shoshone River when that stream flowed through the Pryor Gap (Fig. 6). Accurate surveys prove, however, that the present elevation of the floor of the gap near its center is more than 700 feet higher than the corresponding point on the restored Shoshone profile. The present floor of the gap is more than 500 feet higher than the old Shoshone channel gravels at the point, 18 miles upvalley, where they disappear beneath a thickening mantle of slope wash from the western flanks of the Pryors. This seemingly anomalous relation, however, actually presents no serious difficulty to the capture theory here proposed. The gap floor is made up

FIGURE 1. POWELL AND CODY TERRACES
McCollough Peaks and the dark frontal scarp of the Powell Terrace in the distance. The Cody Terrace is represented by broad flats, interrupted by low scarps in the foreground. (Photo by Douglas Johnson.)

FIGURE 2. POLECAT BENCH FROM THE POWELL TERRACE
Southern marginal scarp of Polecat Bench, as seen from the Shoshone Valley. Westward-dipping sandstones and shales are truncated by the bench surface. (Photo by Douglas Johnson.)

VIEWS IN THE SHOSHONE VALLEY

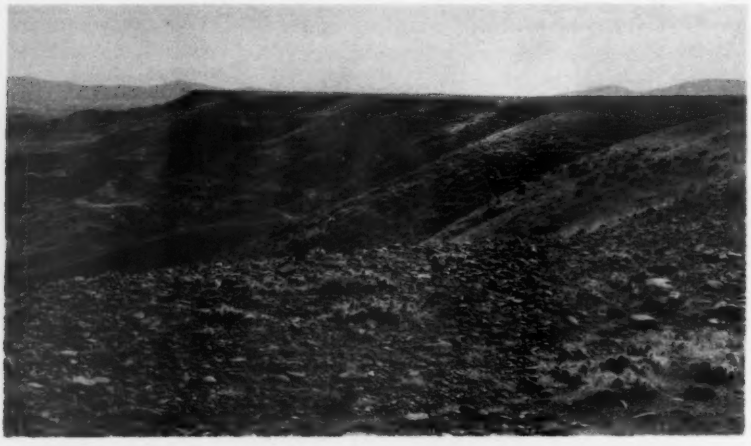


FIGURE 1. SOUTHERN END OF THE BENCH
View looking southwestward. The steep scarp trends diagonally across the old Polecat Shoshone valley floor (*see* Plate 4). The prominent gap on the horizon is the upper open valley of the Shoshone River in the plunging Rattlesnake Mountain arch, at the western side of the Big Horn Basin. (Photo by Douglas Johnson.)

FIGURE 2. SLOPE WASH ALONG THE NORTHERN MARGINAL SCARP OF THE BENCH
The Polecat Shoshone gravel sheet shows as a distinct band across the spurs, resting on inclined sandstone beds. The upper surface of the light-colored wash, which overlies the gravel, slopes gently downward to the left, away from the scarp shown, and toward the axial valley of the bench. The wash was spread over the old valley floor of the Polecat Shoshone River by minor streams heading against the divide between the Polecat Shoshone and the Clark Fork rivers (compare with Plate 7). (Photo by Douglas Johnson.)

of a series of alluvial fans, each extending from a side stream canyon mouth to the opposite valley side or to the foot of an adjacent fan (Pl. 9). In view of the known presence of wash more than 100 feet thick on terraces in present stream valleys in the Big Horn Basin, where the relief is relatively small, it does not seem unreasonable to believe that a maximum of 700 feet of wash may have accumulated in a narrow mountain gorge no longer occupied by any through-flowing stream. Alluviation by side streams is, therefore, tentatively accepted, pending projected geophysical measurements of the depth to rock floor, as the explanation of the abnormal elevation of the gap. An alternate possibility, which cannot be excluded, is that the height of the center of the gap may be due, in part, to arching of the Pryor Mountains since the Polecat Shoshone capture.

Reversal of Sage Creek.—Sage Creek rises in the Pryor Mountains, enters the gap at its southwestern, or upstream, end, and flows south, up the old valley of the Shoshone for a distance of 14 miles before turning southeastward to join the present Shoshone. This peculiar course is interpreted as the direct result of the alluviation of the Pryor Gap. Sage Creek and other tributary drainage from the western slopes of the Pryor Mountains probably continued to flow northeastward through the gap in the period immediately following the diversion of the main stream, at first occupying the abandoned main-stream channel, but soon swinging in a pseudomeandering course around the bases of side-stream fans.[32] Alluviation, because of plentiful waste shed from steep sides into a relatively narrow valley, proceeded more rapidly, and to greater depth, near the center of the gap than near its ends. The original slope of the floor was, consequently, reversed, and Sage Creek was turned back and possibly ponded for a time near its western end. At the present time, the original slope of the abandoned main-stream valley floor has been reversed to the farthest point upvalley reached by waste from the western flanks of the Pryor Mountains, the elevation of the alluvial fill at the center of the gap being 5050 feet, at its southwestern end, 4865 feet, and at the point where the alluvium thins and the old Shoshone gravel appears, 4575 feet. Sage Creek flows up the abandoned main-stream valley (at its far western side, at the foot of a long waste slope from the Pryor Mountains) in a course that is seemingly consequent on the

[32] Shifting of a divide, after capture or glacial diversion, to a side-stream fan in the abandoned main-stream valley has been noted by many workers. O. T. Jones has called the writer's attention to what appears to be the first application of a specific term, "corrom," to such fan-divides. [P. F. Kendall and E. B. Bailey: *Glaciation of East Lothian south of the Carleton Hills*, Roy. Soc. Edinburgh, Tr., vol. 46, pt. 1 (1908) p. 1-33.] A footnote (p. 8) states: "Corrom (cothrom) is a Gaelic word used in place names in the Ardgour district of Argyllshire, to denote a delta watershed. Its literal meaning is a "balance" and it is intended to illustrate that a stream issuing upon such a cone has the chance of flowing either the one way or the other."

valley fill. At Warren, where the waste filling thins, because the course of the old Polecat Shoshone diverges from the flanks of the Pryor Mountains, the drainage of the waste-choked valley found its way into the headwaters of an aggressive present Shoshone tributary, which had already breached the southeast valley side. Here, Sage Creek has cut a moderately wide valley diagonally across the pre-existing Shoshone valley floor, remnants of which form extensive northward-sloping terraces capped with volcanic gravel, in the valley of a small southward-flowing stream (Sage Creek) that carries chiefly limestone.

Pryor Gap landslide.—An irregular hummocky ridge extends almost entirely across the Pryor Gap Valley at its northeastern end. The surface of the transverse ridge is much higher than is the corresponding point on the restored profile of the main stream, which is supposed to have passed through the Gap; hence, if the surface were underlain by bed rock in place, its presence would necessitate drastic revision of all that has been written here in regard to the early history of the Shoshone River. Limestone is exposed at many points on the hummocky surface, but the dip and strike of each of the outcrops differ markedly from those of immediately adjacent outcrops, and at least two of the limestone formations of the nearby valley sides are in close and random juxtaposition. These facts, together with the highly irregular and pitted form of the surface, and the further fact that the bases of some large limestone blocks are exposed in a small valley, cut through the barrier, prove that it is a jumble of landslide debris rather than bed rock. The distribution of the material and the presence of typical landslide scars high on the precipitous north valley side indicate that the slide came from the northwest.

The great size of some of the blocks—the largest seen is approximately 12 by 25 feet and of unknown thickness—and the great aggregate mass of the debris suggest that the slide might have acted as a temporary dam for even so large a stream as the Shoshone. It is considered, however, highly improbable that the occurrence of the slide and the diversion of the stream at a point nearly 50 miles upvalley could have been in any way connected, for the barrier is more than 400 feet lower than the point at which the diversion occurred. The slide probably occurred after the withdrawal of the Shoshone, and probably caused a temporary ponding of Pryor Creek, which has subsequently cut a valley across it.

Extension of Pryor Creek.—Pryor Creek rises in the Pryor Mountains, enters the Pryor Gap near its northeastern end, cuts its way through the landslide just discussed, and follows the abandoned course of the Shoshone to the Yellowstone. Like all the relatively high-gradient small

A sheet of black Polecat Shoshone gravel rests on beveled Paleocene beds and is overlain by a thick mantle of wash, sloping toward the axial valley of the bench. The edge of the gravel sheet marks the northwestern edge of the Polecat Shoshone valley floor. Diagram indicates the proximity of the present northwestern margin of the bench; when the retreating scarp reaches the dot-dash line in the diagram, the condition shown in Plate 6, figure 2, will be duplicated. (Photo by John Lucke.)

NORTH SIDE OF THE POLECAT SHOSHONE VALLEY

FIGURE 1. CONTRASTED STREAM GRAVELS IN PRYOR CREEK VALLEY
Gravel below the inked line is made up largely of basic igneous rock tpyes; overlying gravel is largely limestone. The lower gravel was derived from the distant Absaroka Mountains and was deposited by the Polecat Shoshone River; the upper gravel was derived from the nearby Pryor Mountains and was deposited by Pryor Creek after the diversion of the Polecat Shoshone to the present Shoshone Valley.

FIGURE 2. CLARK FORK — ROCK CREEK DIVIDE
View southward from southern end of Mesa Bench. The smooth surface (right) slopes westward toward Rock Creek, which issues from the Beartooth Mountains in the distance. The westward-retreating scarp faces the headwater basin of Bear Creek, a Clark Fork tributary that will eventually capture Rock Creek (compare with Figure 8). Base of the light band in the scarp marks the contact of bed rock and a thick sheet of gravel and slope wash, derived in part from the former extension of the Mesa Bench to the mountain front. (Photo by J. C. F. Siegfriedt.)

PRYOR CREEK VALLEY AND CLARK FORK — ROCK CREEK DIVIDE

streams that occupied the main-stream valley, Pryor Creek was forced to aggrade markedly, particularly near the mountains. But downcutting of the Yellowstone since the withdrawal of the Shoshone has permitted the Creek to cut considerably below the old Shoshone valley floor, in the lower part of its course; consequently, the Shoshone profile and the Pryor Creek profile cross each other. The photograph (Pl. 8) showing the Pryor Creek limestone gravel resting on the Shoshone volcanic gravel was taken near the limit of Pryor Creek aggradation. Subsequent downcutting of the creek, which involves upvalley shifting on the crossing point of the two profiles, has carried the creek considerably below this site of former upbuilding (Fig. 6). Farther downvalley, where Pryor Creek has cut far below the level of the old Shoshone Valley floor, it has, in downcutting, left minor terraces capped with limestone gravels. The contrast between the 3- to 4-mile-wide terrace capped with volcanic gravel and the much smaller limestone gravel terraces of the inner valley expresses clearly the difference in size and source of the past and the present occupants of the valley.

CLARK FORK-ROCK CREEK AREA

INTRODUCTION

The Rocky Mountain front trends almost directly northward along the western side of the Big Horn Basin to a point, approximately 50 miles north of the Shoshone canyon, where it turns sharply to the west around the great westward extension of the Montana Plains. The Clark Fork River issues from its canyon between the Absaroka and the Beartooth units of the upland about half way between the Shoshone and the angular bend of the mountain front, turns sharply northward, and flows to the Yellowstone River (Pl. 10). Rock Creek issues from the mountains just west of the angle and flows north, parallel with the Clark Fork for approximately 30 miles, then turns and joins that stream just south of the Yellowstone. The Rock Creek valley is relatively shallow and has well-defined terraces from the mountain front to its junction with the Clark Fork. In marked contrast, the Clark Fork Valley is a broadly open lowland with few main-stream terrace remnants. Mesa Bench, on the divide between the two, is the highest feature of the "Plains" adjacent to this section of the mountain front, its south end standing nearly 2,300 feet higher than the Clark Fork lowland and 350 feet higher than Rock Creek.

As in the case of Polecat Bench, the Mesa Bench is at once the major problem and the most important single clue to the erosional history of the area. Thus, it will be shown that before the Mesa stage of regional erosion the relations in size and in declivity of the Clark Fork River and

Rock Creek were probably notably different from those of the present time. A major capture is postulated, which rationally accounts for the change in the relative activity of the two streams and, at the same time, explains certain other striking relations of the present Clark Fork drainage, which are otherwise anomalous. The post-capture history of the two streams is discussed in later sections.

EARLY EROSIONAL HISTORY

Origin of the Mesa Bench.—The Mesa Bench [33] is an elongate tableland, dropping off sharply to the Clark Fork Valley on the east and to that of Rock Creek on the west, and sloping northeastward between them for approximately 20 miles, at a rate that decreases from a maximum of 90 feet per mile near the mountain front to 65 feet per mile at the distal end. The smooth surface truncates bed-rock structures and is everywhere mantled by coarse river gravel. On descending slopes around the bench margins, a talus of resistant gravel commonly covers weaker underlying sandstones, thus giving the gravel sheet an appearance of considerable thickness and obscuring the nature of the rock floor on which it rests; but wherever exposures are fresh, the gravel is found to maintain a remarkably uniform thickness of approximately 18 feet. This uniformity in gravel thickness throughout the length of the bench and the neat beveling of bed-rock structures indicate that the surface on which the gravel rests was formed by the lateral corrasion of the same swinging river (or rivers) that deposited the gravel—that is, the Mesa Bench is a surface of stream planation.

The position of the bench on the Clark Fork—Rock Creek divided suggests that it might have been formed by either, or both, of these streams; the essential identity in composition of the Mesa gravels with those on lower terraces in the Rock Creek valley, and the marked differences between these gravels and those of undoubted Clark Fork terraces (Pl. 3, symbols 4 and 6) prove that the bench was formed by Rock Creek. But the present eastern margin of the bench is an abrupt scarp, varying from 500 to 1500 feet in height, which descends to the Clark Fork lowland. Manifestly, no such feature could have been present when Rock Creek was cutting laterally at the Mesa level, for the stream would certainly have sooner or later spilled over the edge.

Many minor valleys, trending northwestward across the Mesa surface toward Rock Creek, are truncated at the present eastern edge of the bench (*note* Clear Creek, Fig. 8). As these valleys are cut distinctly below the

[33] The terminology adopted here for the Rock Creek terraces and benches is that used in Guidebook 24, *Yellowstone-Beartooth-Big Horn region*, Intern. Geol. Congr., 16th ses. (1933). The writer's interpretation of the Mesa Bench differs from that advanced in the Guidebook, where it is stated (p. 39) that "The varying thickness of the coarse torrential deposits coating the Mesa peneplain indicates that this graded surface was partly made by the filling of valleys."

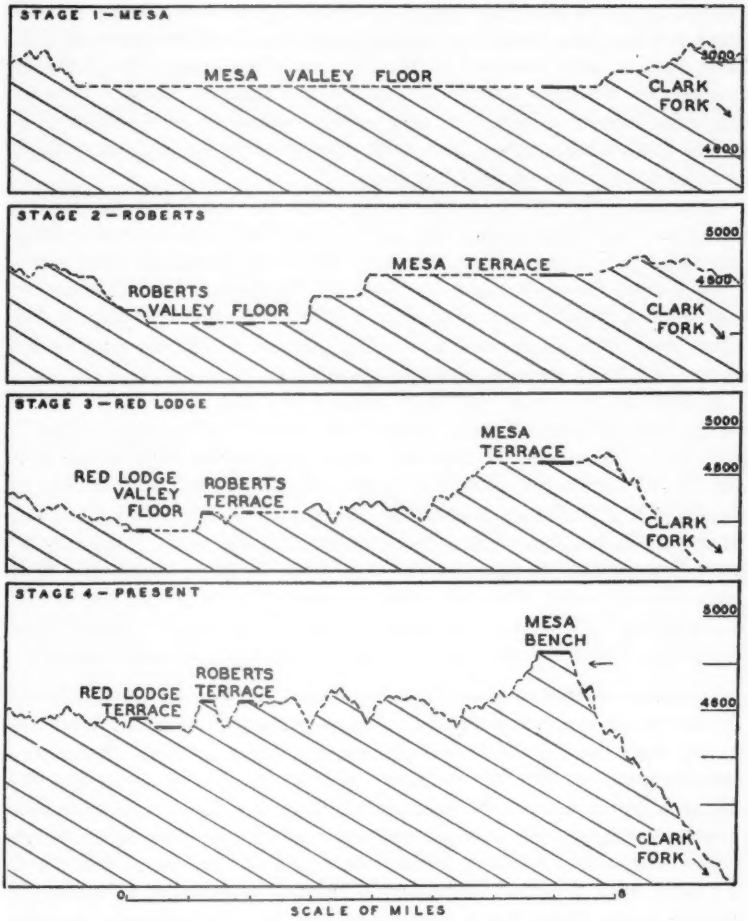


FIGURE 7.—*Evolution of cross-profile of Rock Creek valley*

Heavy black dashes represent remnants of Rock Creek gravel sheets; all other parts of the profiles are hypothetical (stages 1, 2, and 3), or approximate (stage 4). The profiles are drawn along an arc of a circle described from the Rock Creek canyon mouth and crossing the present valley at Roberts, approximately 12 miles from the mountain front.

bench surface to its very margin, it seems certain that they formerly extended farther to the east, and have been beheaded as a result of westward retreat of the scarp. Similar truncation of roads, necessitating long detours around fresh landslide bites along the bench margin, indicates that

westward retreat is progressing rapidly at the present time. In one small section, two landslides, involving several hundred acres of the Mesa surface, have occurred within the past seven years. Small ponds now occupy the inner margin of some of the slumped blocks. It is of particular interest to note that each of these slides drops a portion of a former Rock Creek valley floor, still mantled with Rock Creek gravel, down into the valley of the Clark Fork.

Commonly, the gravel-mantled surface of the bench extends to the top of the descending scarp, but, near its southern end, a narrow strip underlain by bed rock stands distinctly above the bench level and separates it from the steep slope to the Clark Fork (Fig. 8). This higher land defines, locally, the eastern limit of Rock Creek swinging at the Mesa level; it is interpreted as the last remnant of the old east valley side of the Mesa stage Rock Creek valley floor.

It appears, then, that the Mesa surface was first formed as a broad valley floor of Rock Creek, limited by normal valley sides (Fig. 7, stage 1). Tributaries entered the valley from higher land to the east and flowed across the floodplain to the stream. Later downcutting of Rock Creek transformed part of the old valley floor into a broad terrace, similar to the terraces of the present valley, and, like them, limited by an ascending stream-cut scarp on the east (Fig. 7, stages 2 and 3). Still later, this old valley side was almost completely destroyed by the retreat of the Clark Fork slope; a complete reversal of topography resulted, and the terrace was transformed into the present Mesa Bench (Fig. 7, stage 4).

Imminent capture of Rock Creek.—The difference in elevation between the low-gradient Clark Fork and the high-gradient Rock Creek increases to the south as the distance from the accordant junction of the two streams increases (Pl. 10, profiles). The pushing back of the present divide between the two (a direct result of their difference in elevation) increases progressively in vigor and has reached progressively more advanced stages from north to south, with the result that the present divide is nearer Rock Creek as the mountain front is approached. The smooth surface of the Mesa Bench narrows to a point approximately one mile north of the mountain front and is completely cut away for the remainder of the distance (Fig. 8). The bench is already explained as a former Rock Creek terrace, transformed into a divide bench through the destruction of former higher land to the east. Where the bench itself has been destroyed the present divide is on the next lower Rock Creek terrace. Wash on this lower planed surface sloping toward Rock Creek from the very edge of the break strongly suggests the former presence of higher land where now a steep scarp drops off toward the Clark Fork drainage; the presence in the wash of pebbles identical with those of the Mesa

Bench gravel sheet indicates that the higher land was made up, in part, at least, of the former continuation of the Mesa surface toward the mountain front (Pl. 8, fig. 2).

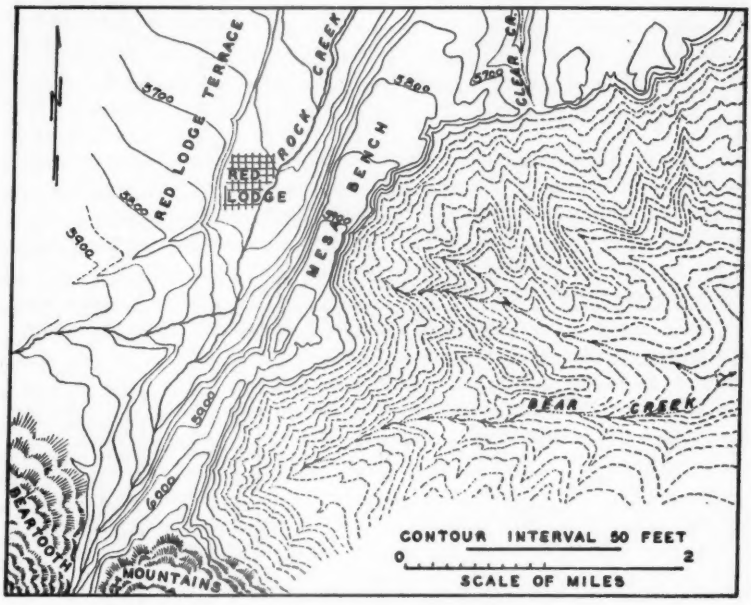


Figure 8.—*Topographic sketch map of the Red Lodge area*

The divide between Rock Creek and the headwaters of Bear Creek is within half a mile of the former stream and is shifting westward. Clear Creek is one of many Rock Creek tributaries that have been beheaded in the process. The hills east of Clear Creek stand distinctly above the level of the Mesa Bench and represent part of the former divide between the Mesa stage valley floor of Rock Creek and the Clark Fork Valley. Dashed contours in the Bear Creek drainage are approximate, intended only to show the asymmetry of the present divide. (Compare with Plates 10 and 11.)

As the lower terrace that now forms the divide near the mountain front is the first major terrace above the present floodplain of Rock Creek, it is evident that continued westward retreat of the divide will soon lead to the diversion of that stream. How very imminent the capture is may be seen from the character of the present divide (Pl. 8, fig. 2; Fig. 8). The headwaters of Bear Creek, the Clark Fork tributary that will effect the capture, are only one mile from Rock Creek and approximately 300 feet below the Rock Creek channel. The presence of perennial springs in the Bear Creek valley suggests that a covert capture is already in operation along the bedding planes of porous Fort Union sandstones, which strike across the divide.

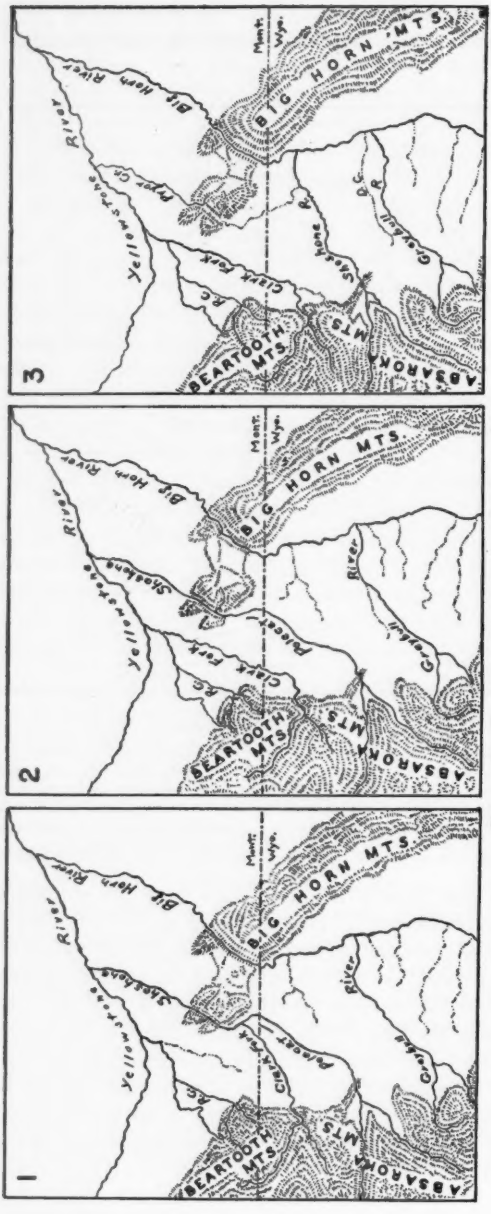


FIGURE 9.—*Stream captures in the Big Horn Basin*

Maps showing evolution of the drainage of the northern part of the Big Horn Basin, according to the hypothesis involving capture of the Clark Fork from the Polecat Shoshone River. Stage 1, stream pattern before the Clark Fork capture (?); stage 2, stream pattern before the Polecat Shoshone and Greybull captures; stage 3, present pattern.

CAPTURE OF THE CLARK FORK RIVER

Anomalous relations of Clark Fork and Rock Creek. — The transformation of a part of the Mesa stage valley floor of Rock Creek into the present Mesa Bench is due, as already shown, to a vigorous westward shifting of the Clark Fork - Rock Creek divide. This shifting indicates that the divide is decidedly out of adjustment to the relative elevations of Clark Fork and Rock Creek. But divides in an "n plus one" cycle of erosion are commonly well adjusted to the elevations of the adjacent streams, for the condition of equilibrium,

established in the earlier cycle or cycles, is inherited in later cycles. The lack of adjustment of the Clark Fork-Rock Creek divide, therefore, suggests that some marked change has recently occurred in the relative gradients of the two streams, such that, at any considerable distance from their accordant junctions, the elevation of the Clark Fork has been greatly decreased in relation to that of Rock Creek. One working hypothesis that is in harmony with some of the pertinent considerations may be stated briefly.

Capture hypothesis.—According to this hypothesis, Rock Creek formerly flowed northward to the Yellowstone as a separate stream. The stream issuing from the present Clark Fork Valley, between the Beartooth and the Absaroka mountains, then flowed eastward to join the pre-Polecat Shoshone River in the Big Horn Basin west of Pryor Gap, the combined volume of the two streams flowing through the gap and to the Yellowstone in the valley now occupied by Pryor Creek (Fig. 9, stage 1). Some time before the Polecat stage, a minor tributary of Rock Creek (or of the Yellowstone), working headward from the north, tapped what was then the Clark Fork of the Shoshone and led it off to its present course to the Yellowstone (Fig. 9, stage 2). This introduction of the large Clark Fork River into a valley formerly occupied by a much smaller stream caused vigorous downcutting and initiated the westward shifting of the Clark Fork-Rock Creek divide, which has transformed a former Rock Creek terrace into a divide bench, and which will soon culminate in the capture of Rock Creek. The Shoshone River, weakened by the loss of the Clark Fork, continued to maintain its course through the Pryor Gap until the Polecat stage,[84] when it was tapped and led off by a Big Horn tributary (Fig. 9, stage 3).

Implications of capture hypothesis.—(1) It has been pointed out in an earlier section that the present asymmetrical position of the Big Horn River in the Big Horn Basin may be rationally explained by direct superposition from a waste fill, and by this method alone. Similarly, the former (Polecat) course of the Shoshone, trending northeastward across the basin to the Pryor Mountains and thence northward for many miles along the very foot of that range, is in complete accord with the former drainage pattern reconstructed by deduction, for the Pryor Mountains are low and presumably furnished little of the basin-filling waste. The present course of the Clark Fork, however, is seemingly anomalous, for that stream turns sharply northward at its canyon mouth and flows nearly

[84] The theory that the Clark Fork River formerly joined the Polecat Shoshone in the basin west of Pryor Gap and had been captured by a Yellowstone tributary before the capture of the Polecat Shoshone is advanced in Guidebook 24, *Yellowstone-Beartooth-Big Horn region*, Intern. Geol. Congr., 16th ses. (1933) p. 34.

parallel to, and only a few miles distant from, the high and rugged Beartooth Mountains, which constitute the greatest potential source of waste around the borders of the basin. The former course implied by the capture theory here under examination is, on the other hand, completely in harmony with the pattern that might be expected in the filled basin, both the Clark Fork and the Shoshone rivers being pressed against, and later superimposed across, the nose of the Pryor Mountains, by waste derived from high ranges to the west (Fig. 9, stage 1).

(2) If the Clark Fork did formerly flow eastward through the Pryor Gap, its present course to the Yellowstone implies that some condition, or conditions, favored capture by a small Yellowstone tributary. It will be shown in a later section that all the major streams of the basin region are, by their nature, notably vulnerable to capture by small streams, even when the small streams have none of the advantages commonly recognized as causes of capture. In the case under consideration, the supposed captor was favored by a shorter course to the master Yellowstone and by its position on the strike of weak rocks. A general southward shifting of divides between eastward-flowing streams throughout the Montana Plains, to be discussed in connection with the problem of the age of the terraces and benches of the Big Horn region, would have favored the capture. The vigorous shifting of the present Clark Fork-Shoshone divide toward the latter stream, indicated by the asymmetry of the divide east and west of Heart Mountain and by the truncation of Shoshone tributaries at the divide, may be due, in part, to this general southward shifting of divides and, in part, to rejuvenation of the Clark Fork by capture.

(3) The diagnostic character of the gravel now transported by the Clark Fork has been mentioned in connection with the origin of Polecat Bench; the absence of granitic (Clark Fork) pebbles in the Polecat gravels and in the correlative deposits in the Pryor Creek valley was stated as one line of evidence supporting the thesis that the Clark Fork did not join the Shoshone at the Polecat stage. One of the most obvious implications of the hypothesis that it did join the pre-Polecat Shoshone is that, if such had been the case, granitic pebbles should be found in the basin between the canyon mouth and the Pryor Gap, or mixed with Shoshone gravel on pre-Polecat terraces in the Pryor Creek valley, provided, of course, that an appreciable amount of granite was available for transportation by the Clark Fork branch of the Shoshone River.

No gravel of any kind has been found between the Clark Fork canyon and the Pryor Gap. This circumstance can hardly be considered unfavorable to the capture theory, however, for, from a vantage point on low morainal hills at the canyon mouth, it is possible to look directly

FIGURE 1. AIR VIEW OF PRYOR GAP

Looking southwestward, diagonally across the valley cut by the Polecat Shoshone River through the Pryor Mountains. Alluvium, deposited by side streams after the withdrawal of the Polecat Shoshone, is believed to be approximately 700 feet thick near the center of the gap. (Photo by Charles Belden.)

FIGURE 2. PRYOR GAP "CORROM"

Alluvial fans, built out from side-stream canyons, form the divide between northeastward- and southwestward-flowing drainage in the abandoned Polecat Shoshone Valley. (Photo by Douglas Johnson.)

across the maturely dissected central part of the basin to the Pryor Gap. If, as suggested here, the Clark Fork formerly followed that course, its bed must certainly have been some hundreds of feet higher than the line of sight; the stream gravels, if they were ever present, have been carried away in the deep erosion that has produced the present surface.

Northeast of the Pryor Mountains, however, pre-Polecat gravels are preserved over wide areas on the divide between the Pryor Creek (old Shoshone) valley and that of the Big Horn River, and between the Pryor Creek and the Yellowstone valleys, but no granitic pebbles have been found in such a position that it may be stated with certainty that they passed through Pryor Gap. This fact presents a serious obstacle to acceptance of the capture theory. That it may not necessarily be fatal to the theory is indicated by the following considerations. The Clark Fork Valley in the mountains forms the boundary between the Absaroka Range, composed of Tertiary lavas and agglomerate, on the south, and the Beartooth Range, on the north (Pl. 3). The Beartooth Range consists of a pre-Cambrian core, still partly covered, particularly on the southern flanks, by Paleozoic sedimentary rocks. The granitic pebbles now transported by the Clark Fork are derived from a narrow inner gorge, cut in crystalline rocks, and from those parts of the core laid bare by removal of, first, the Absaroka volcanics, which almost certainly formerly extended north of the canyon and, second, the Paleozoic strata. No direct evidence is at hand to indicate the level at which the Clark Fork was flowing immediately before the supposed capture, but the Polecat stage valley floor of the Shoshone, necessarily later than any practicable Clark Fork capture, stands now 800 feet higher than the level of the Shoshone River and 625 feet (Fig. 6) higher than the level of the Yellowstone. If downcutting of the Clark Fork River has been proportionate to the downcutting of the Shoshone, the pre-capture level of the Clark Fork may have been considerably above the level of the open valley above its present gorge (approximately 1200 feet above present stream level). This open valley marks the approximate level of the contact between the crystalline rocks and the overlying sedimentary rocks, and the relations between these two rock types and the volcanics in typical north-south structure sections across the valley indicate that it is entirely possible that little, or no, crystalline waste was available for transportation by the Clark Fork at this high pre-capture level.

Conclusion.—The evidence indicating the passage of the Polecat-Shoshone River through the Pryor Gap is direct, and seems as nearly conclusive as geologic proof may be. In sharp contrast, the evidence for the passage of the Clark Fork through the gap during some pre-

Polecat stage is indirect and may be regarded as suggestive but not conclusive. Future capture of Rock Creek by the Clark Fork, and, at a much later time, of the Shoshone by the Clark Fork, may be predicted with reasonable assurance, if the works of man are disregarded. The problem of disentangling the drainage history of the area after these captures have occurred will be an interesting one, indeed.

TERRACES OF ROCK CREEK VALLEY

The Mesa Bench represents the highest, and oldest, Rock Creek valley floor of which any trace remains; successively lower terraces of the valley proper are gravel-capped surfaces, representing successively later stages of stream-cutting. The highest, or Roberts, terrace and the next lower, or Red Lodge, terrace stand at different levels on opposite sides of the valley, and the gravel sheet of each is essentially flat in cross-valley section; the two terraces, therefore, indicate two distinct pauses in the downcutting of Rock Creek (Fig. 7, stages 2 and 3). The present valley floor differs from all the higher corrasion surfaces in that it has been formed, in part, at least, by aggradation.

The terraces of the first 30-mile stretch of the Rock Creek valley northward from the mountain front are essentially continuous, and the terrace gravel rests on surfaces that bevel more or less pronounced structures in Cretaceous and early Tertiary strata. For the remainder of the distance to the Yellowstone River, high-level gravel sheets are preserved only on small isolated buttes or mesas, underlain, for the most part, by massive, nearly horizontal sandstones of Cretaceous age. The correlation of these latter, discontinuous remnants is complicated by uncertainty in each case as to whether the present surface is the original surface of deposition or whether the gravel has been let down to a level controlled by the resistant sandstone beds. Terrace remnants were not mapped in this region.

The data bearing on the longitudinal slopes of the Rock Creek terrace and bench sequence are presented graphically, without verbal description of individual remnants, in the accompanying longitudinal profiles (Pl. 10, profiles). The profiles were constructed by projection of the terrace surfaces to a vertical plane through the axis of the present valley floor, each point being projected along an arc of a circle described from the mouth of the Rock Creek Canyon as a center. According to the argument advanced in discussion of the genetically similar terraces of the Shoshone Valley, each of the profiles represents a former Rock Creek valley floor, possibly altered in declivity by differential movement subsequent to its formation.

The Mesa Bench and the Roberts Terrace slope less steeply than the present stream; the Red Lodge Terrace slopes more steeply than the

stream or the higher terraces; and a buried valley floor, indicated by thickness of alluvium in a few deep wells near Red Lodge, probably slopes less steeply than the stream. It should be emphasized that, although these differences in declivity are slight in terms of feet per mile, they are, nevertheless, far greater than the limits of error in the surveying on which the terrace profiles are based. Interpretation of the variations in slope of the successive valley floors, in terms of regional geomorphic history, may be deferred until all the pertinent evidence is before the reader.

CLARK FORK VALLEY

General statement.—As has already been pointed out, the Clark Fork Valley differs from all the other main stream valleys of the Big Horn Basin in that few main-stream terrace remnants are preserved. Several isolated, gravel-capped buttes stand 300 feet higher than the stream, approximately 9 miles northeast of the canyon mouth, and are probably to be correlated with several other high floodplain remnants, which can be traced for some 20 miles downvalley. Much farther downstream, terraces are more or less continuous between the junction of Rock Creek and the point where the Clark Fork enters the Yellowstone, but, because these two terraced areas are separated by a long stretch of valley where none is well preserved, correlation is uncertain, and no attempt is made at restoration of any former river profile. The present valley floor is aggraded, but, here again, the paucity of reliable data as to the depth of valley fill makes determination of the slope of the buried river bed impossible. For these reasons, little information bearing on the problem of regional geomorphic history has been derived from study of the Clark Fork, but the valley is of particular interest; first, because of a remarkable development of broad lateral corrasion planes formed by minor streams (discussed in a separate section) and, second, because of a striking difference in composition between the terrace gravels and those of the present floodplain.

Contrasted gravel types in Clark Fork Valley.—The areal extent of outcrop of distinctive rock types in the drainage basin of the Clark Fork indicates that the gravel transported by the stream at the mountain front might be expected to consist of approximately 35 per cent volcanic pebbles, 55 per cent crystalline pebbles, and 10 per cent sedimentary pebbles. This estimate is closely in accord with pebble counts made on isolated gravel-capped buttes northeast of the canyon mouth and on related terrace surfaces (Pl. 3, symbol 6). It is most decidedly out of accord with counts made on the gravel of the present floodplain, where commonly less than five per cent of volcanic pebbles are found (Pl. 3, symbol 7). The difference in appearance between the Clark Fork ter-

race gravels, rich in dark basaltic rock types, and the dominantly crystalline gravels of the immediately adjacent valley floor is quite as striking as the difference between either of these types of Clark Fork gravel and the gravel of any of the other major streams of the basin. The floodplain is continuous upvalley to a point where it merges with (or is overspread by) an outwash plain, heading at a moraine that partly closes the canyon mouth; the same abnormal scarcity of volcanic pebbles characterizes the moraine.

The notable difference in the two types of Clark Fork gravel indicates some distinct difference in the source from which they were derived. The presence of deeply cut cirques and other marks of profound glacial erosion in the Beartooth Mountains, north of the Clark Fork Canyon, suggests that this range was more severely glaciated than were adjacent parts of the Absaroka Mountains to the south, and that the predominance of crystalline pebbles in the glacial outwash might have been due to a flooding of the Clark Fork Valley by granite-carrying Beartooth ice. This hypothesis is most satisfactorily verified by Dake, in a brief paper in which he calls attention to the anomalous presence of moraines composed largely of crystalline pebbles in the valleys of southern tributaries of the Clark Fork, which drain areas where only volcanic rocks are available for erosion. Dake states:—

"In the light of these facts it would seem that the Beartooth Plateau was occupied by an ice cap that sent several tongues eastward into the edge of the Bighorn Basin, and a number southward across the Clarks Fork Canyon several miles *up the valleys* of Sunlight, Elk, Russell and Dead Indian Creeks."[35]

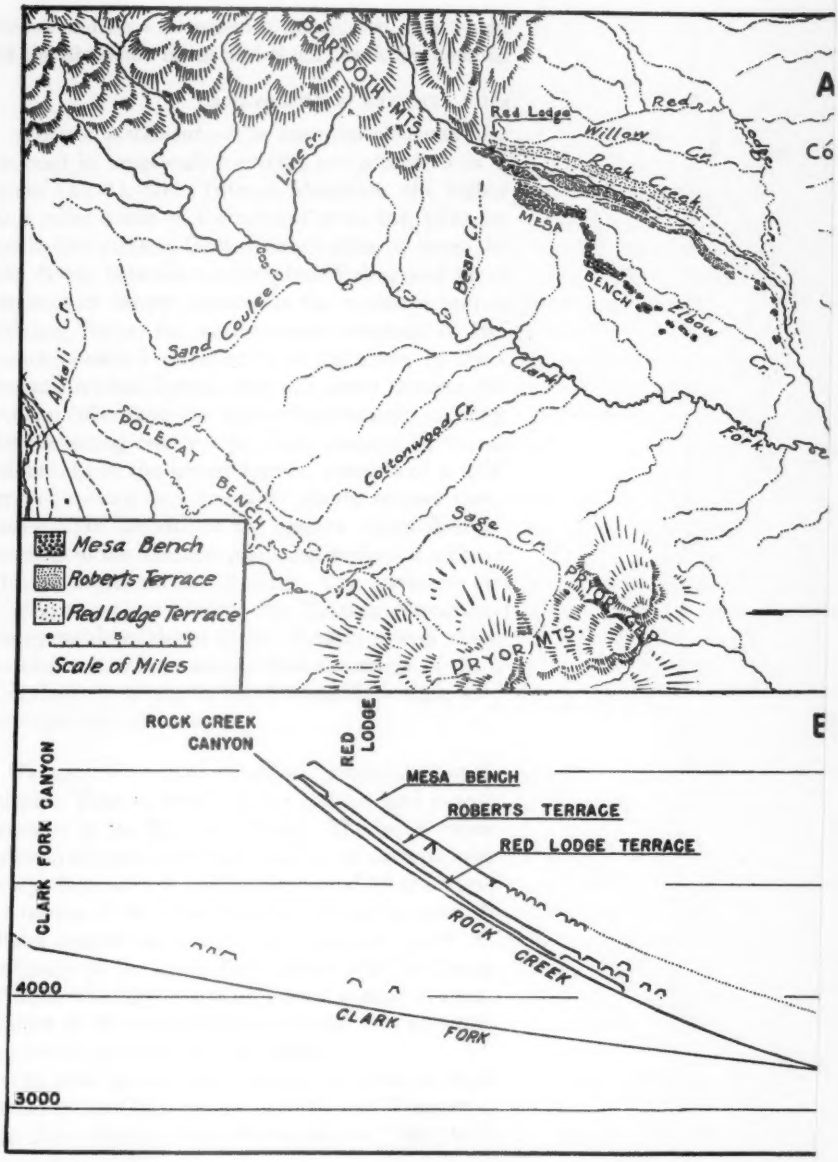
The relation of the terraces and benches of the Big Horn Basin to periods of glaciation in the mountains around the basin is of primary importance in any interpretation of the regional significance of these features. In this connection, the striking difference in composition between the fluvial gravels of Clark Fork terraces and the fluvio-glacial outwash gravels of the aggraded valley floor will be discussed in a later section.

GREYBULL RIVER AREA

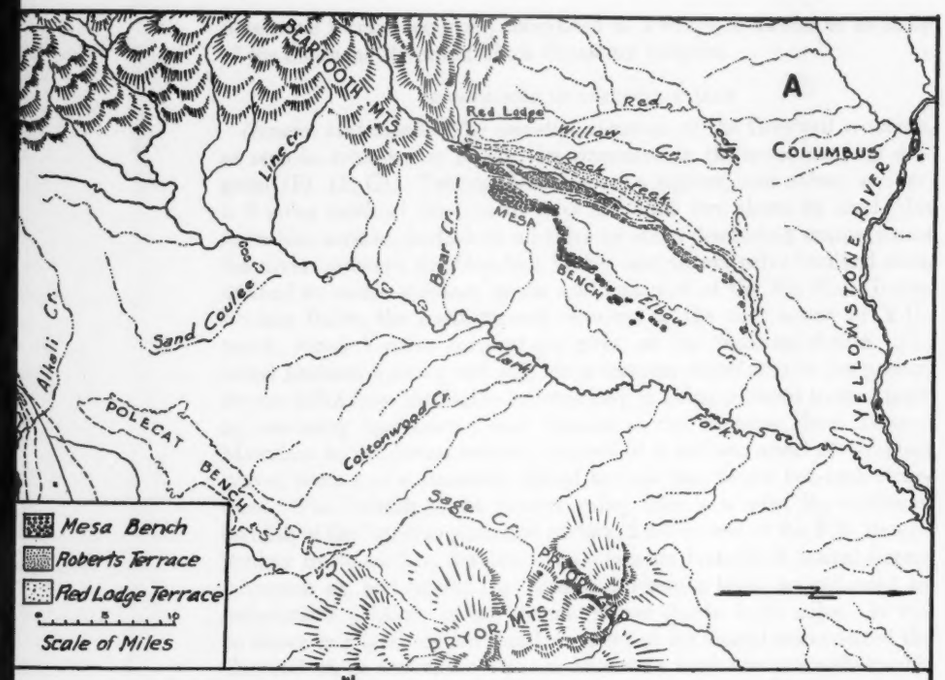
INTRODUCTION

Polecat Bench, in the Shoshone area, and the Mesa Bench, in the Clark Fork-Rock Creek area, are, in a sense, special or exceptional forms, directly related to a series of relatively recent drainage changes. The broad, gravel-capped surfaces now to be described differ from these in this important respect, that they represent the normal, uninterrupted activity of only one major stream, the Greybull River. The Greybull

[35] C. L. Dake: *Glacial features on the south side of Beartooth Plateau, Wyoming*, Jour. Geol., vol. 27 (1919) p. 131.



CLARK FORK–ROCK CREEK AREA



CLARK FORK–ROCK CREEK AREA

bench and terrace sequence, considered as a whole, is as simple as those of the Shoshone River and Rock Creek are complex.

CROSS-PROFILE OF GREYBULL VALLEY

General statement.—The essential elements of the Greybull sequence, as seen in cross-valley profile, are presented in the accompanying diagram (Pl. 11, C). Tatman Mountain, the highest, and oldest, surface, is 8 miles south of the present river and 1230 feet above its level. Its table-like surface, limited on all sides by steep descending scarps, forms the divide between the Greybull Valley and an extensive badland area, drained by minor streams, in the southern part of the Big Horn Basin. Bridger Butte, the northernmost remnant of the next lower, or Y.U., bench, stands 7 miles north of the river, on the Shoshone divide. The broad Emblem Terrace and still lower terraces descending to the present stream differ from the higher benches only in being confined to the valley by ascending rear scarps; each element in the sequence, from Tatman Mountain to the lowest terrace, consists of a uniform sheet of Greybull gravel, resting on a smoothly planed surface that bevels bed-rock structures. The breadth of the present valley floor is a mile; the minimum breadth of the Emblem planation surface, 2 miles; and of the Y.U. Bench-Bridger Butte surface, 6 miles. The minimum breadth of lateral stream migration at, and below, the Tatman Mountain level, as indicated by indisputable evidence of the channel gravel sheets, is 14 miles. It will be shown in a later section that the extent of the lateral migration of the Greybull at, or above, the Tatman Mountain level was probably double or triple this distance.

Tatman Mountain.—Tatman "Mountain", or, for purposes of this discussion, Tatman Bench, is the highest, and oldest, gravel-capped erosion remnant in the Big Horn Basin. The bench extends for approximately 8 miles in an east-west direction, but its smooth, eastward-sloping surface is everywhere narrow and is interrupted by three prominent notches, formed by retreat of the lateral scarps. From its crest, the superb panorama includes almost the whole of the desolate waste of fantastically dissected badlands of the basin floor, interrupted by the perfectly smooth slopes of numerous high-standing lateral corrasion remnants and by the green valleys of the present main streams, and surrounded on all sides, except the north, by snow-capped ranges.

The crest gravel sheet, which contains no fossils, was tentatively assigned to the Oligocene by Sinclair and Granger, on the basis of its position above middle or late Eocene shales.[86] But the planed surface on which

[86] W. J. Sinclair and Walter Granger: *Eocene and Oligocene of the Wind River and Big Horn basins*, Am. Mus. Nat. Hist., Bull., vol. 30, art. 7 (1911) p. 88, 111.

the gravel rests truncates structures of the shales at a low angle, which varies from point to point along the bench margin. The bench gravel is composed largely of basic igneous rock types, carried by the Greybull River from the Absaroka Mountains, although no pebbles of comparable lithology have been found in gravel beds interstratified with the shales.[87] Evidently, then, the lapse of time represented by the disconformity between the two was sufficient for the accumulation of the great pile of Absaroka lavas and agglomerates.

The shales just beneath the planed surface on which the gravel rests are so highly carbonaceous as to have encouraged searchers for coal to dig numerous prospect pits; these beds, or similar beds at a slightly lower level, have yielded the flora likened by Berry to that of the South Atlantic coastal States (p. 818). No geologist, standing on the outcrop of the leaf-bearing lignitic shales, could fail to be impressed by the striking contrast between the indicated humid climate of early Tertiary time and the present conditions of advanced aridity; a rational explanation of the climatic change has already been suggested. But the overlying bench gravels are identical in structure and in thickness with the gravels that mantle the several lower benches and terraces descending to the present Greybull River, and the eastward slope of the bench gravel sheet is essentially the same as that of the valley floor now in process of formation by the river; that is, the bench gravel "belongs" in the present arid floor of the basin. The disconformity at the base represents the long period, during which the climatic change occurred.

The interpretation proposed here differs from that suggested by Sinclair and Granger in that the Tatman gravels are believed to have been deposited, not by aggrading streams, during, or at, the culmination of the early Tertiary period of basin filling, but by the degrading Greybull River, during the current period of re-excavation.

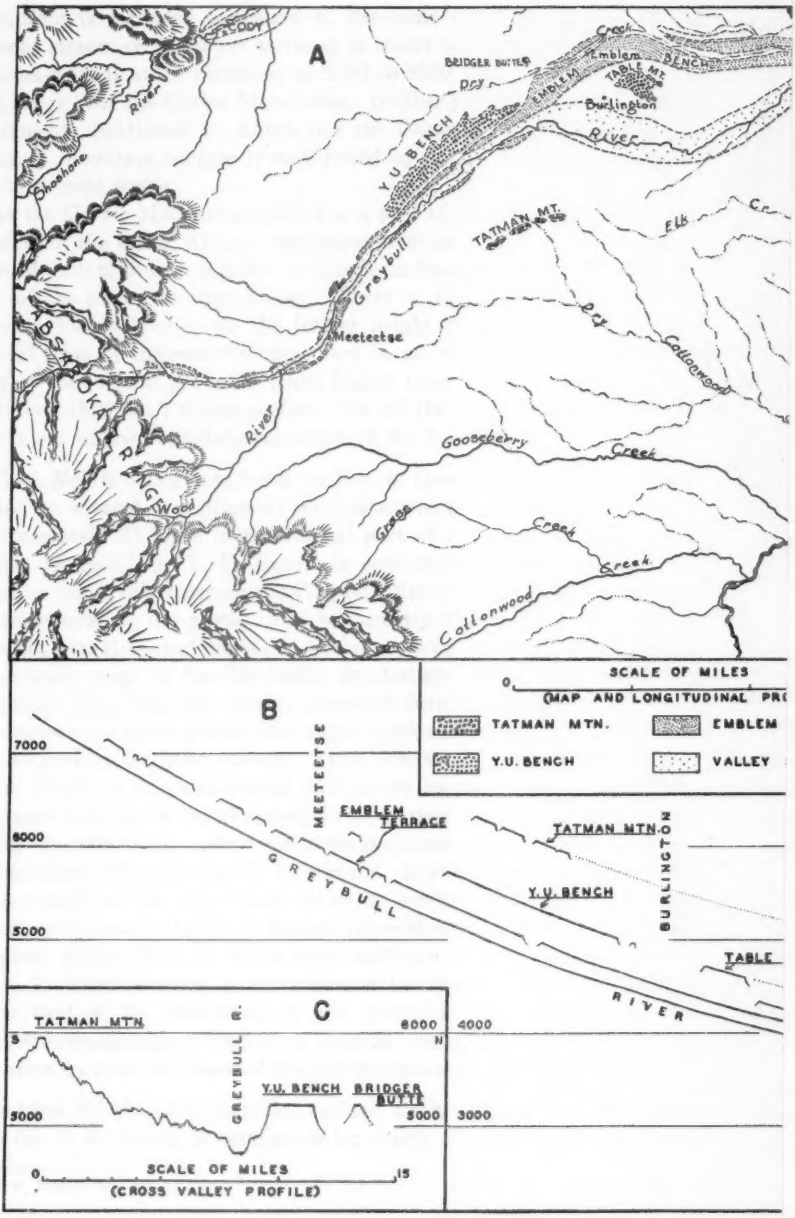
Alden recognized Tatman Mountain as a remnant of a former Greybull corrasion plane, but correlated it with high spurs of the Absaroka Mountains:

> "A stream gradient projected eastward 30 to 35 miles from the edge of the smoothly worn top of the Carter Mountains, 10,000 feet above sea level, would coincide with the gravel cap of Tatman Mountain, if an initial slope of 250 feet to the mile is supposed to decrease gradually to a fall of about 40 feet to the mile."[88]

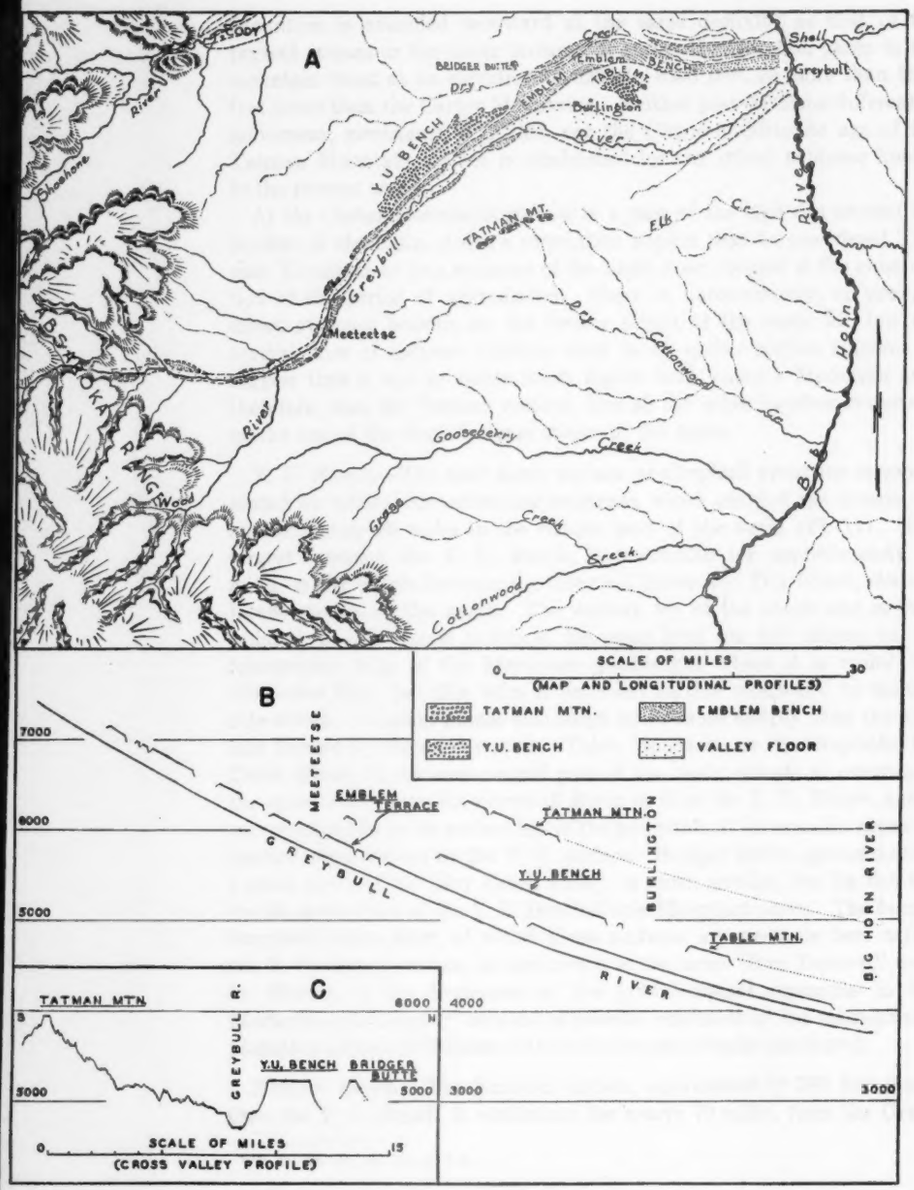
The assumed declivity of 250 feet per mile is more than four times as steep as the present Greybull profile or the profile of any Greybull terrace in the section of the valley under consideration; if the profile of Tatman

[87] D. F. Hewett: *Geology and oil and coal resources of the Oregon Basin, Meeteetse and Grass Creek Basin quadrangles, Wyoming*, U. S. Geol. Surv., Prof. Pap. 145 (1926) p. 30-47.

[88] W. C. Alden: *Physiography and glacial geology of eastern Montana and adjacent areas*, U. S. Geol. Surv., Prof. Pap. 174 (1932) p. 11.



GREYBULL–DRY CREEK AREA



GREYBULL–DRY CREEK AREA

Mountain is extended westward at the same declivity as that of the present stream or the lower terraces, it enters the Greybull gorge in the mountain front at an elevation of 8200 to 8500 feet, or more than 1500 feet lower than the Carter Mountains. Neither post-Miocene differential movement, mentioned by Alden, nor the Oligocene-Miocene age of the Tatman Mountain surface is established by any direct evidence known to the present writer.

As the Carter Mountains surface is a part of the high rim around the borders of the basin, Alden's correlation implies that he considered Tatman Mountain to be a remnant of the basin floor, formed at the culmination of the period of aggradation. There is, unfortunately, no positive direct evidence bearing on the former height of the basin fill, but the several lines of indirect evidence cited in an earlier section combine to suggest that it was probably much higher than Tatman Mountain and, therefore, that the Tatman surface, like all the other benches examined, marks one of the degradational stages of the basin.

Y. U. Bench.—The next lower surface of Greybull planation is represented by several discontinuous remnants, which parallel the stream for approximately 35 miles in the central part of the basin (Pl. 11). The largest remnant, the Y. U. Bench, is continuous for approximately 15 miles on the divide between the Greybull River and Dry Creek, the adjacent stream on the north. The western tip of the bench and several isolated, gravel-capped buttes at the same level are well shown on the topographic map of the Meeteetse quadrangle, where it is called the Meeteetse Rim, but this term is extended farther westward, to include side-stream corrasion planes that slope much more steeply than the surface formed by the main stream. Table Mountain, on the Greybull-Dry Creek divide, in the east-central part of the basin, stands at essentially the same level above the Greybull River as does the Y. U. Bench, and a surveyed profile on its surface lies in the projection of the smooth, concave-upward slope defined by the Y. U. surface. Bridger Butte, approximately 3 miles north of the Dry Creek valley, is much smaller, but its flat top stands at the level of the Y. U. Bench-Table Mountain curve. The former Greybull valley floor, of which these surfaces are parts, is here called the Y. U. Bench surface, in preference to the term "Rim Terrace," used by Hewett in his discussion of the gravel-capped remnants in the Meeteetse quadrangle,[39] because of possible confusion of the main-stream planation surface with those of the side streams already mentioned.

Emblem Bench.—The Emblem surface, approximately 360 feet lower than the Y. U. Bench, is continuous for nearly 70 miles, from the Grey-

[39] D. F. Hewett: *op. cit.*, p. 5-8.

bull canyon mouth to the Big Horn junction. In the western half of the basin the remnants of the old Emblem valley floor are present as normal river terraces on one or both sides of the present Greybull Valley, but near the center of the basin the gravel-mantled surface, maintaining its smooth eastward slope, crosses a low divide and continues to the Big Horn River, in the valley now occupied by Dry Creek. The terrace is not represented in the present Greybull Valley east of the cross-over point.

It is clear from these relations that, at the Emblem stage, the Greybull River flowed to the Big Horn through what is now the eastern part of the Dry Creek valley, and that it was diverted to its present valley by capture at, or near, the cross-over point. Dry Creek, formerly a minor tributary just below the point of capture, has continued to occupy the valley segment abandoned by the main stream. The reason for the capture has been treated at length in another article,[40] and may be stated here briefly. The large Greybull River was completely graded, from the point of capture to the Big Horn River. The capturing stream, draining only a small part of the arid basin floor, could have carried water for only a short time each year, and it had none of the advantages, such as shorter distance to base level or weak channel rocks, commonly found in the diversion of large streams by smaller streams. The Greybull was high and vulnerable to capture because its graded declivity, adjusted to the transportation of a great load of coarse rock waste from the Absaroka Mountains, was actually steeper than the declivity of the much-smaller capturing stream, which transported only fine sands and silts derived from the friable Tertiary beds of the basin floor. The capture, as an event in the Emblem stage cutting of the Greybull River, is of minor importance insofar as regional geomorphic history is concerned, but the principle involved is, as will be shown in a later section, of critical importance in the interpretation of drainage changes that may have occurred earlier, during the degradation of the basin.

Stream-cut, gravel-mantled terraces are developed at several levels between the Emblem surface and the present floodplain, and some moderately extensive remnants, intermediate between the Y.U. and the Emblem surfaces, are preserved in the eastern part of the basin, but none is traceable for any considerable distance. The present stream is fully graded from the mountain front to its junction with the Big Horn.

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[40] J. H. Mackin: *The capture of the Greybull River*, Am. Jour. Sci., 5th ser., vol. 31 (1936) p. 373-385.

CORRELATION AND AGE OF TERRACES AND BENCHES

INTERVALLEY TERRACE CORRELATION

It is evident that absolute elevation is no reliable guide in the correlation of sloping terraces in different valleys. The common divergence of terrace profiles from the profiles of present Big Horn Basin streams makes long-range correlation on the alternative basis of similarity in elevation above the stream unsafe even in a single valley; it is doubly fallible when used in correlation from valley to valley. Degree of dissection and degree of weathering of gravels vary so widely even on the same terrace in the same valley that these criteria could not be used in inter-valley correlation. Correlation of corresponding members of similar sequences of terraces in adjacent valleys is seemingly valid, provided the sequences exhibit the same distinctive features with sufficient clarity to establish the essential identity of the histories of the two valleys. The single method that may be used with complete confidence in precise regional correlation of stream terraces is an extension of that used in earlier sections of this paper in correlation of remnants in a single valley—namely, the tracing of terraces in two, or more, valleys to the point where they join at the valley junctions, or to points where terraces in two, or more, valleys join the same terrace in the valley of a master stream.

Several circumstances combine to make the satisfactory application of this ideal method impossible in the valleys of the Big Horn Basin area. Two "master" streams, the Yellowstone and the Big Horn rivers are involved. Tracing of terraces in the master valleys between tributary junction points is complicated in the case of the Yellowstone by difficulty in distinguishing true stream-cut terraces from cappings of river gravel held up by the nearly horizontal sandstones in which the Yellowstone Valley is cut, and, in the case of the Big Horn, by the 30-mile gorge incised by that stream in the resistant rocks of the Big Horn Mountain arch. Moreover, none of the higher terraces and benches in the valleys of the basin are continuous to junctions with master-valley terraces.

The data on which conclusions as to the relative ages of the stream-planed surfaces of the basin may be based are assembled in graphic form on the accompanying chart (Fig. 10). The position of the short dash before the name of each of the several planation levels is determined by projection of the longitudinal profile of each surface to a vertical plane cutting the axis of the master-stream valley. The projection of the longitudinal profile of Polecat Bench and the correlative terrace in the Pryor Creek valley, for instance, stands 625 feet above the present level of the Yellowstone at Huntley and represents the approximate level above the present Yellowstone at which the old Polecat-Shoshone River joined that stream (Fig. 6). Possible variations in the levels of the former junctions,

introduced by uncertainty as to the declivity of the projection and by doubt as to the former geographic position of the axis of the master valley, are expressed by figures indicating possible errors plus or minus, which

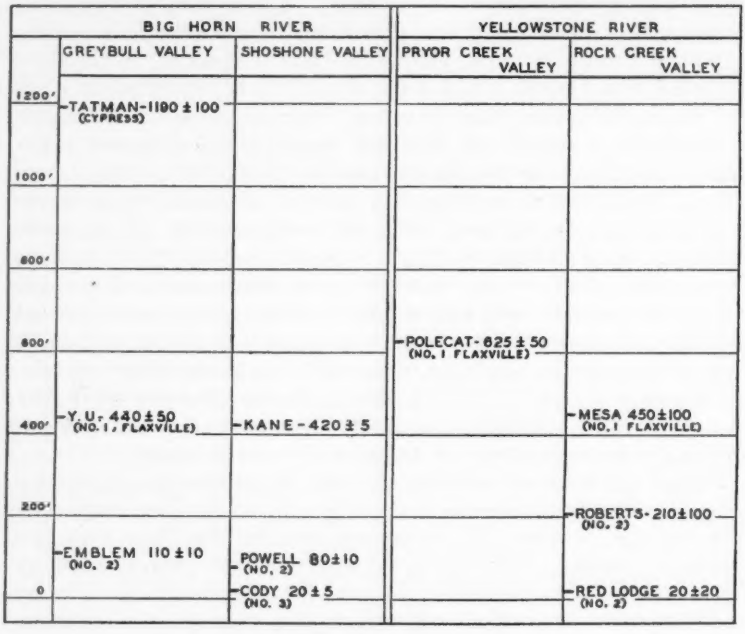

FIGURE 10.—*Big Horn Basin terrace levels*

Chart shows relative heights of the projected profiles of the terraces and benches of the Big Horn Basin above the present levels of the Yellowstone and Big Horn rivers. Names and numbers in brackets indicate correlations assigned by W. C. Alden.

follow the elevation figures. These possible errors are large where possible variations in the location of the high-level junctions are considerable and where lengths of projection, compared with lengths of known profile, are large, as in the case of Tatman Mountain or the Mesa Bench (Pls. 10 and 11). The figures and names in brackets indicate the correlations assigned by Alden and adopted in numerous United States Geological Survey reports.[41]

The Cody Terrace in the Shoshone Valley (Big Horn system, Fig. 10) and the Red Lodge Terrace in the Rock Creek valley (Yellowstone system) stand at, or near, the present valley floor levels at the mouths of

[41] W. C. Alden: *op. cit.*, pl. 1.

the two streams and diverge markedly from the floodplains of the streams as they are traced upvalley. Tentative correlation is based on the assumption that the present valley floors of the Big Horn and the Yellowstone rivers opposite the tributary mouths represent the same stage and that the strikingly similar terraces in the tributary valleys represent the last stage except one.

The Powell and the Emblem terraces, in the Greybull and the Shoshone valleys, respectively, indicate pauses in the downcutting of the two streams, interrupted, in the case of the Greybull, by capture and, in the case of the Shoshone, by quickened downcutting. Tentative correlation of the terraces, as representing a distinct stage of regional erosion, is based on the fact that the projections of both terraces stand at approximately the same height above the present level of the Big Horn River, at points only 20 miles apart. The two terraces are probably not exactly contemporaneous, for, as already pointed out, the Greybull was withdrawn from the Emblem level by capture. Alden's correlation of the Roberts Terrace in the Rock Creek valley with the Powell and the Emblem terraces is supported by the similarity in sequence in the several valleys, each being characterized by a well-defined pause in downcutting (Powell, Emblem, and Roberts terraces), followed by a stage of somewhat quickened downcutting, in which a moderately broad valley floor was formed (Red Lodge and Cody terraces and unnamed, discontinuous remnants in the Greybull Valley), and this, in turn, by relatively rapid downcutting in the mountainward parts of the valleys to (or below) the present floodplain levels. The difference between the levels of the projections of the Powell and the Emblem terraces above the Big Horn and that of the Roberts Terrace above the Yellowstone may be due to changes in the declivities of the Yellowstone or the Big Horn, to the retarding effect of the hard rocks of the Big Horn Mountain arch on post-Emblem-Powell downcutting of the Big Horn, or to warping. Discontinuous terrace remnants, standing 300 feet above the level of the Clark Fork near its canyon mouth and decreasing in elevation downvalley, are probably equivalent to the Roberts-Emblem-Powell stage in the adjacent valleys, as indicated by Alden.

The former Polecat Shoshone River joined the Yellowstone approximately 30 miles downvalley from the mouth of Rock Creek, and the projection of the Polecat Bench profile of the old Shoshone is 175 feet higher above the present level of the Yellowstone than is the projection of the Mesa Bench profile of Rock Creek. If it is supposed that the Polecat and the Mesa surfaces are identical in age, the difference in the heights of their projected profiles above the Yellowstone must be ascribed to markedly greater incision of the Yellowstone opposite the mouth of the

Polecat Shoshone than opposite the mouth of Rock Creek since the surfaces were formed; this difference in depth of incision might be due to warping, of the type postulated by Alden in discussion of the Yellowstone terraces. An alternate, and simpler, explanation of the difference in height of the projected tributary profiles is that the Polecat surface was developed at a higher level and is somewhat older than the Mesa surface.

In either instance, the Clark Fork River must have occupied its present valley during the formation of the Mesa Bench by Rock Creek, for, as has been shown in an earlier section, any practicable diversion of the Clark Fork, from a former course across the Basin to its present northward-trending valley, must have occurred before the Shoshone had reached the Polecat stage; in other words, as Polecat Bench is the same age as, or older than, the Mesa Bench, and as the present course of the Clark Fork is older than Polecat Bench, then the present course of the Clark Fork must be older than the Mesa Bench.

Even if, as already suggested, Polecat Bench is somewhat older than the Mesa Bench, it does not follow that the two surfaces represent two distinct stages of regional degradation, for the Shoshone was withdrawn from its Polecat valley by capture, which may have occurred during, but some time before the close of, the stage of slow downcutting in which the Mesa Bench was formed by Rock Creek. The highest terrace in the present Shoshone Valley (the Kane terrace) stands at approximately the same height above the stream near its mouth as does the projection of the Mesa surface above the Yellowstone at the mouth of Rock Creek; it is believed to be a remnant of a broad valley floor cut by the Shoshone River soon after its diversion to the Big Horn, and to be essentially equivalent in age to the Mesa Bench surface in the Rock Creek system and to the corresponding Y.U. surface in the Greybull system.

Tatman Mountain is by far the highest, and, consequently, the oldest, of the stream-planed surfaces of the basin floor. The height of its projected slope above the level of the Big Horn River (1125 feet) supports Alden's correlation of the Tatman surface with gravel-mantled surfaces standing 1000 to 1200 feet above the level of the Yellowstone on Pine Ridge, in the acute angle between the converging Yellowstone and the Big Horn rivers.[42]

Thus, the stream-cut surfaces of the Big Horn Basin fall, in a general way, into the five stages of regional erosion recognized by earlier workers on the Montana Plains, the first being the Tatman-Pine Ridge stage (Cypress, of Alden), the second the Y.U.-Mesa (-Polecat ?) stage (No. 1, Flaxville, of Alden), the third the Roberts-Emblem-Powell stage (No.

[42] *See also* W. T. Thom, Jr., G. M. Hall, C. H. Wegemann, and G. F. Moulton: *Geology of Big Horn County and the Crow Indian Reservation, Montana*, U. S. Geol. Surv., Bull. 856 (1935) p. 66-69, pl. 3.

2, of Alden), the fourth the Red Lodge-Cody stage (No. 3, of Alden), and the fifth the present stage. It should be emphasized that discussion of the several surfaces in the Big Horn Basin in terms of the stages recognized by Alden is based on Alden's provisional correlation from the type localities of the several surfaces to the Big Horn region, and does not imply that the present writer accepts as final the identity in age of the Mesa Bench, for instance, with the Flaxville Plain of northern Montana. Neither does this general correlation imply that any, or all, of the surfaces assigned to the same stage in the Big Horn region were precisely contemporaneous; it is recognized that quickening of stream cutting may work upstream slowly, particularly in hard-rock belts, and that minor impulses of rejuvenation may become confused or lost in the process. Intermediate terrace remnants are present in each of the valleys studied. The tentative correlations are advanced to establish a general conception of local age relations and to stress differences (as in the case of Polecat and Mesa benches) as well as general similarities in age; the single important conclusion (that the downcutting of the streams has been pulsational) that might be considered to be based on the correlations, is attested by the terrace sequence in each individual valley.

RELATIONS OF TERRACES AND BENCHES TO GLACIAL STAGES

Two, and possibly three, glacial stages have been recognized in the mountains bordering the Big Horn Basin.[43] Moraines of the last, presumably Wisconsin, stage stand near the present river levels at, or near, the mountain front in Rock Creek, Clark Fork, and Greybull valleys, and approximately 100 feet higher than the stream in the valley of the South Fork of the Shoshone River, 20 miles west of Rattlesnake Mountain.[44] An earlier moraine has been described at the level of the Roberts Terrace in the Rock Creek canyon mouth.[45] These relations raise a question of fundamental importance, bearing on the origin of the higher gravel-capped surfaces of the basin—namely, whether the capping gravels were deposited by heavily loaded streams issuing from ice tongues during glacial stages, or whether the terraces have no direct genetic relation to glaciation, but are interglacial or preglacial.

Alden,[46] who has mapped terraces and continental- and valley-glacier moraines in Montana and northern Wyoming; Bevan,[47] who has studied

[43] Arthur Bevan: *Glaciation northeast of Yellowstone National Park* [abstract], Geol. Soc. Am., Bull., vol. 42 (1931) p. 325-326.

[44] J. T. Rouse: *The physiography and glacial geology of the Valley region, Park County, Wyoming*, Jour. Geol., vol. 42 (1934) p. 738-752.

[45] Arthur Bevan and Erling Dorf: *Yellowstone-Beartooth-Big Horn Region*, Intern. Geol. Congr., 16th ses., Guidebook 24 (1933) p. 44.

[46] W. C. Alden: *Physiography and glacial geology of eastern Montana and adjacent areas*, U. S. Geol. Surv., Prof. Pap. 174 (1932) 133 pages.

[47] Arthur Bevan: *Rocky Mountain peneplains northeast of Yellowstone Park*, Jour. Geol., vol. 33 (1925) p. 563-587.

the Beartooth Mountain moraines; and Rouse,[48] who has described moraines in the valley of the South Fork of the Shoshone, agree that terraces of the type treated here are not genetically associated with valley moraines. Relations observed by the present writer in the valley of the South Fork of the Shoshone and in the Greybull Valley at the mountain front indicate that certain of the terraces pass beneath the moraines and are distinctly older, but in neither place is it clear whether or not the terraces now mantled by moraine were trenched by stream action before the moraines were formed. No terraces cut in moraine have been observed.

Two additional lines of evidence bear directly on the problem. The terraces and benches of the Big Horn Basin are remnants of broad valley floors cut in bed rock by long-continued stream action at the same, or nearly the same, level. The great breadth of the old planation surfaces indicates a remarkable stability in the delicate balance of load and discharge of the streams, which is not to be expected during glacial stages. The contrast between the stream-planed, gravel-veneered surfaces of the terraces and benches and the deeply aggraded present valley floors associated with the last ice advance suggests that the terraces do not represent earlier glacial stages.

It has been shown, moreover, that, in the case of the Clark Fork, the debris washed out from the last valley glacier differs markedly in composition from the normal fluvial gravel of that stream. It is reasonable to suppose that the distribution of ice and the relative intensity of glaciation in the Clark Fork drainage basin were essentially the same during possible earlier glacial stages as during the last stage, and, therefore, that the detritus washed out by the stream during the earlier glacial stages should have been as distinctively rich in crystalline rock types and as poor in basaltic types as was that washed out from the last glacier. As the gravel that mantles the surfaces of the high-level terraces and buttes in the Clark Fork Valley is of "normal" composition, markedly different from the glacial outwash, it appears that these surfaces, which are typical of terraces and benches associated with other main streams of the basin, were not formed during earlier glacial stages, but must be preglacial or interglacial.

AGE

No fossils have been found in terrace gravels in the Big Horn Basin or on high terraces in the Yellowstone Valley, to the north. The age of the stream-planed surfaces of the basin must, therefore, be determined by comparison with features of known age in adjacent regions. Alden has shown, by relations of terraces with advances of continental ice sheets

[48] J. T. Rouse: *op. cit.*, p. 738-752.

in the upper Missouri drainage, that his No. 2 surface is Pleistocene,[49] and this age assignment may be accepted for the No. 2 correlatives in the Big Horn region. The problem of the ages of higher, supposedly pre-Pleistocene, surfaces is closely associated with the general problem of the date of initiation of downcutting in the present cycle; this problem necessitates discussion of an apparent conflict between evidence gathered in the Montana Plains, on the one hand, and in the Wyoming Basin and adjacent regions to the south, on the other.

An Oligocene fauna has been discovered in fluvial gravels that underlie the Cypress Hills Plateau, in southwestern Saskatchewan.[50] Alden considers the plateau surface to be Oligocene, and he tentatively correlates these gravels with the gravel-capping on Pine Ridge, between the Yellowstone and the Big Horn rivers, and, in turn, with the Tatman Mountain gravel sheet in the Big Horn Basin. His provisional correlation of the Tatman surface with the subdued surface at 10,000 feet on Carter Mountain, at the western margin of the basin, indicates, as mentioned in an earlier section, that he considers the date of the beginning of the present cycle of degradation to be late Oligocene or Miocene.[51]

Gidley has assigned to the late Miocene or early Pliocene a fauna collected by Collier and Thom from the Flaxville gravels, which cover broad upland areas north of the Missouri River in Montana.[52] Alden considers the Flaxville surface to be post-Cypress in age and correlates with it the Mesa Bench and other high-level surfaces in the Big Horn Basin, holding that these represent an early stage of regional degradation.

Grading of the High Plains depositional surface in Colorado, presumably completed before the present cycle of downcutting began, continued into Pliocene time. Blackwelder describes the "Sherman peneplain" of the Laramie Mountains and other ranges of southern and central Wyoming as a surface produced by the aggradation of the plains and basins and degradation of the mountain masses; his placing of the Sherman stage in the Pliocene and all later degradational stages in the Pleistocene has been generally, but not universally, accepted.[53]

In harmony with this relatively late date for topographic forms in southern regions of Wyoming is Bevan's conclusion that the "Sub-summit peneplain" of the Absaroka and the Beartooth ranges is Pliocene.[54] The

[49] W. C. Alden: *op. cit.*, p. 31-58.

[50] L. M. Lambie: *The vertebrata of the Oligocene of the Cypress Hills, Saskatchewan*, Canada Geol. Surv., Contr. Pal., vol. 3, pt. 4 (1908) p. 7.

[51] W. C. Alden: *op. cit.*, p. 42.

[52] A. J. Collier and W. T. Thom, Jr.: *The Flaxville gravel and its relation to other terrace gravels of the northern Great Plains*, U. S. Geol. Surv., Prof. Pap. 108 (1918) p. 179-184.

[53] Eliot Blackwelder: *Post-Cretaceous history of the mountains of central western Wyoming*, Jour. Geol., vol. 23 (1915) p. 206.

[54] Arthur Bevan: *Rocky Mountain peneplains northeast of Yellowstone Park*, Jour. Geol., vol. 33 (1925) p. 563-587.

dating of the sub-summit surface is based on Bevan's belief that the surface bevels the Absaroka volcanics, thought to be Miocene in part, and on other evidence; Rouse, in the most recent physiographic study of the Absaroka Mountains, accepts Bevan's age assignment.[55] As has been pointed out in an earlier section, the Carter Mountain surface, which bevels the volcanics near the Absaroka Mountain front, is probably decidedly older than the Tatman surface of the basin. If, as seemingly indicated by relations to the south and in the Absaroka Range, the sub-summit surface is Pliocene, the Tatman surface must be Pliocene or younger, as opposed to its assignment to the Oligocene or early Miocene on the basis of Alden's correlation southward from the Cypress Hills. No complete understanding of the Tertiary history of the Big Horn Basin region is possible until these anomalous dates are reconciled, and, as the critical localities are widely scattered beyond the limits of the Big Horn Basin, and constitute, in themselves, several major problems, the best that can be attempted here is an analysis of the problem and a brief discussion of alternative possibilities.

The fossiliferous gravels of the Cypress Hills are said to vary in thickness from 50 to 500 feet.[56] Statements of thickness of bench gravels, based on observations made incidentally in the course of other geological work, are to be accepted with reservation, for the apparent thickness may be due, in large part, to a talus of resistant gravel descending over weak underlying rocks (Pl. 1), but descriptions of the exposures around the margins of the Cypress Plateau admit no doubt as to the correctness of the figures. The great and variable thickness of the Cypress gravel contrasts sharply with the moderate and uniform depth of gravel on benches and terraces of the Big Horn Basin and raises a question as to the validity of the fossils as indices to the age of the topographic surface of the plateau.

Two possibilities immediately suggest themselves: first, the Cypress gravels might have been deposited by aggrading streams on an earlier-prepared erosion surface, the present top of the plateau representing the highest level of upbuilding; or, second, aggradation might have continued to a higher level, and the present surface gravels might have been formed by laterally swinging streams, during a distinctly later degradational stage. In the first instance, the Cypress surface would be essentially similar to the High Plains and to the surface of Tatman Mountain as interpreted by Sinclair and Granger; in the second instance, it would correspond to the interpretation of Tatman Mountain proposed in this paper. Sinclair and Granger suggested that the Tatman surface may be

[55] J. T. Rouse: *op. cit.*, p. 738-752.

[56] R. G. McConnell: *Report on the Cypress Hills, Wood Mountain and adjacent country*, Canada Geol. Surv., Ann. Rept., vol. 1 (1885) 30c-32c.

Oligocene, because the crest gravel sheet rests on fossiliferous Eocene beds. But, as has been shown in an earlier section, the Tatman gravel sheet represents, not continued deposition following formation of the fossil-bearing strata, but a much later degradational stage, and it may be stated with some assurance that the disconformity at the base of the gravel sheet, on which this conclusion is based, in part, would not have been so readily discovered, in any one exposure, had the underlying beds been coarse gravels rather than laminated shales. It is impossible to determine, from the literature, from what exact horizon in the thick Cypress gravels the fossils were taken; it is apparent that in neither of the alternative cases mentioned would the fossils serve to fix the date of the surface of the Plateau, unless they were enclosed in surface layers.

Observations made during one week of field study in the Cypress Hills, with these possibilities in mind, indicate that (1) the gravel sheet that caps the plateau surface and its numerous outlying erosion remnants is relatively uniform in thickness, not exceeding 40 feet; (2) this gravel sheet rests disconformably on sedimentary beds of different lithology and age in different parts of the district; (3) it is not fossiliferous; and (4) it is separated from a thick gravel sequence, which contains numerous bone fragments identified as Oligocene, by the unconformity mentioned in (2).

These relations indicate that there are two distinct gravel beds in the Cypress district—an older sequence, containing an Oligocene fauna, for which the name Cypress formation should be retained, and an overlying gravel sheet, containing no fossils, which is younger than the Oligocene gravels on which it rests and older than the oldest drift of the Keewatin glacier. It appears, therefore, that the assignment of the topographic surface of the Cypress Hills to the Oligocene on the basis of paleontologic evidence may be in error, and that, insofar as the dating of landforms of the Northern Plains depends on the Cypress surface, some revision may be required.

The paleontologic evidence bearing on the age of the Flaxville surface seems somewhat more compelling, for the fossils were collected from surface gravels, and Collier and Thom present evidence tending to prove that the bone fragments and teeth were not reworked from earlier deposits. Gidley accompanied his identification of the fauna as a whole as late Miocene or early Pliocene with mention of one tooth (Camelops) of probable Pleistocene age, but there is some doubt as to whether or not the tooth really belongs to the Flaxville assemblage.

Identification and dating of topographic forms along the Rocky Mountain front in Colorado is a controversial topic now under discussion in the geological literature. The generally accepted Pliocene date for the initiation of the present cycle in that region is supported by the presence of Pliocene vertebrate remains in the fluvial deposits that form the High Plains surface. It appears, therefore, that, if the assignment of the Flaxville degradational surface of the Northern Plains to the late Miocene

or early Pliocene is correct (and if the higher Cypress surface is somewhat older, even though it may not be as old as Oligocene), a distinct difference in the date of beginning of the present cycle of degradation in the Central and Northern Plains is indicated. Such a difference is suggested independently by the widespread preservation of the old constructional surface in the High Plains surface and its complete, or almost complete, removal in the Northern Plains. The divide between the northward-flowing Powder, Belle Fourche, and Cheyenne rivers and the eastward-flowing North Platte and Niobrara rivers is markedly asymmetrical and is being shifted southward at the present time. The difference in the date of inception of degradation in the Central and the Northern Plains might have been due to a difference in the date of uplift of the two regions, or, as will be suggested on a later page, to a difference in climatic conditions in the two latitudes. In either instance, a wave of rejuvenation would have worked southward, and it is possible that the apparently conflicting testimony from the northern and the southern areas may not actually conflict at all, but that active degradation in Montana in the Pliocene may have been contemporaneous with continued upbuilding or planation in southern Wyoming and Colorado.

According to this hypothesis, the present cycle of downcutting would have been initiated in the Big Horn Basin at a later time than in northern Montana and at an earlier time than in southeastern Wyoming and Colorado; middle or late Miocene would seem an appropriate date. The Tatman surface, representing an early degradational stage, would then be referable to the early or middle Pliocene, the Y.U.-Mesa (Polecat) surface to the late Pliocene, and all lower surfaces to the Pleistocene.

It is evident that the hypothetical erosional history of the Great Plains as outlined depends largely on the identification of the Flaxville gravels as late Miocene or early Pliocene. Alden states:

> "The relations of the oldest drift of the mountain glaciers on the supposed remnants of the Flaxville Plain close to the mountain front lead to the inference that the Flaxville Plain had not been greatly dissected when climatic conditions became such as to bring on extensive glaciation in the mountains."[57]

This type of evidence is difficult to evaluate in terms of age, but, taken together with the shallow depth of stream-trenching between the Flaxville surface and lower terraces of known Pleistocene age, and the relative narrowness of the valley floors opened out in the weak rocks of the Plains by vigorous streams issuing from the mountains in post-Flaxville time, it does establish the alternative possibility that the Flaxville stage is late Pliocene or early Pleistocene. This hypothesis has the advantage of simplicity, in that the degradational stages of the Central and the Northern

[57] W. C. Alden: *Physiography and glacial geology of eastern Montana and adjacent areas*, U. S. Geol. Surv., Prof. Pap. 174 (1932) p. 31.

Plains would be essentially contemporaneous; it is not ruled out by the difference in stage of erosional development of the two areas, for this difference may be due to a difference in the rate of erosion in the two latitudes, rather than to a difference in the date of the beginning of downcutting.

These two alternative possibilities as to the dates of topographic forms in the Northern Plains make up a question of primary importance, involving not only the Tertiary history of the Northern and Central Plains, but the age relations of geomorphic stages in the Rocky Mountains as well. The solution will depend, in large part, on a critical re-examination of the field relations of the fossils by which the gravels of the Northern Plains have been dated. Until such studies have been made, the age of degradational forms of the Big Horn Basin cannot, in the opinion of the writer, be definitely fixed.

INTERPRETATION OF TERRACES AND BENCHES

INTRODUCTION

Great monoclinal flexures and high-angle, range-front faults with measurable displacements of thousands of feet testify to the magnitude of the Laramide differential movements by which the larger structural features of the Big Horn region were produced. Sharp folds in early Tertiary strata around the borders of the basin prove that differential upward movement of the ranges, relative to the basin floor, continued during the period of aggradation that accompanied the major orogeny, but, in the absence of post-Eocene sedimentary beds, similar structural evidence of middle or late Tertiary movement is absent. The tectonic relations of the mountains and the plains during the current cycle of degradation constitute a problem of major regional importance; its solution is to be sought in study of landforms of the plains adjacent to the mountain front.

Richard M. Field, after preliminary surveys made in connection with the research program of the Big Horn-Beartooth-Yellowstone Project, was impressed by the striking similarity between the smoothly sloping, gravel-capped surfaces of the Big Horn Basin and certain forms described and figured by Douglas Johnson in a presentation of the lateral corrasion theory of origin of planed rock floors in desert regions. The present study was undertaken to test the application of the theory of lateral planation in the study of the landforms of the Big Horn Basin, and to determine, if possible, the bearing of these forms on the problem of recent land movements.

PEDIMENT CONCEPT

The extensive development of planed rock floors around the margins of desert ranges of the Southwest has drawn increasing attention in recent

years, and two contrasted theories as to the origin of these features have been proposed; one, that they are produced by the lateral planation of heavily laden streams issuing from the mountain fronts; the other, that they are formed by weathering, sheetwash, and related processes. The several lines of evidence set forth at length in foregoing sections of this paper seem to demonstrate that the high-level, gravel-mantled rock surfaces of the Big Horn Basin must be ascribed to lateral cutting by the present major streams during earlier stages of regional degradation. For this reason, it is desirable, before proceeding to systematic discussion of these forms, to outline briefly the history of the lateral planation concept, in order that the local forms of the basin may be viewed in their proper relation to those described in southwestern regions. The papers of Lawson,[58] and others who advocate weathering as a dominant process in the production of desert planes, have been summarized by Davis, himself a supporter of the weathering theory.[59]

LATERAL PLANATION THEORY OF PEDIMENT FORMATION

The importance of lateral stream corrasion in the fashioning of desert forms appears to have been first recognized by G. K. Gilbert,[60] who, in 1877, ascribed to this process the production of smoothly sloping, gravel-mantled rock planes on the flanks of the Henry Mountains uplift, in Utah. The Henry Mountains surfaces truncate rock structures and are widely opened in weak rock belts crossed by the streams, but they are notably constricted in more resistant rocks; they were regarded by Gilbert as being entirely analogous to the valley floors cut by graded streams in humid regions, and their much higher declivities were explained as a function of the relatively greater load of waste carried by the streams issuing from the arid mountains.

In 1903, Douglas Johnson, following Gilbert, attributed to lateral stream corrasion extensive surfaces, beveling rocks of markedly different degrees of resistance, which slope outward from the Cerrillos Hills and other mountain groups in New Mexico.[61]

In 1912, Sidney Paige described rock planes in the Silver City quadrangle, New Mexico, and explained the surfaces as the result of "interstream erosion", limited in depth by the edge of the sheet of detritus that slopes outward from the range front to the central playa of a closed basin.[62] Paige shows that upbuilding of the central playa should cause

[58] A. C. Lawson: *Epigene profiles of the desert*, Univ. Calif., Pub. Geol., vol. 9 (1915) p. 23-48.
[59] W. M. Davis: *Rock floors in arid and humid climates*, Jour. Geol., vol. 38 (1930) p. 1-27, 136-158.
[60] G. K. Gilbert: *Report on the geology of the Henry Mountains*, U. S. Geog. Geol. Surv., Rocky Mtn. Reg., Washington (1877) 160 pages.
[61] Douglas Johnson: *Geology of the Cerrillos Hills, New Mexico*, Columbia Univ. School Mines Quart., vol. 24 (1903) p. 173-246, 303-350, 456-500; vol. 25 (1904) p. 69-98.
[62] Sidney Paige: *Rock cut surfaces in the desert ranges*, Jour. Geol., vol. 20 (1912) p. 442-450.

the waste apron to overlap the rock plane as it is extended toward the center of the range, and that, at any advanced stage of range destruction, the overlapping sheet should thicken wedgelike from the mountain base toward the center of the basin. It is clear from the context that by "interstream erosion" is meant the lowering of interfluves by weathering and other essentially local processes, but a significant contribution to the lateral planation concept is the recognition of an abrupt break in slope between the relatively steep mountain front and the more gently sloping plane at its base. Paige considers this break to be due to a "second process"—namely, the trimming back of the mountain spurs by the lateral cutting of streams.[63] The cutting is specifically restricted to the edges of alluvial fans, and no mention is made of the earlier papers of Gilbert and Johnson, in which lateral planation is given a major rôle in the beveling of planes apparently identical with those treated by Paige.

Kirk Bryan's excellent descriptions of the landforms of the desert ranges and basins of the Papago country, in southwestern Arizona, published in 1923, provide a wealth of data bearing on the form of rock planes in arid regions, and on the relations of thickness and slope of mantling alluvial deposits to the slopes of the rock floors.[64] The term pediment, first applied by McGee to a specific rock plane in the same area, is used by Bryan as a general term for all similar features. Weathering and rill cutting, controlled in depth by a slope of stream transportation, seem to be regarded as the essential processes causing recession of the mountain front and extension of the rock pediment. The angular break between the steep mountain front and the pediment is explained (following Lawson) as a function of a difference in size of weathered materials on the contrasted slopes, and to an abrupt concentration of rill cutting at the mountain base, but lateral corrasion by streams is mentioned in connection with the beveling of the plane and the recession of the range front, and is placed first in a summary of processes active in the modeling of desert forms. The stream courses of the Papago country slope to the sea, and the streams are, as a rule, degrading. The planed rock floors near the ranges are concave upward in longitudinal profile and are mantled by a thin veneer of alluvium, but well records are introduced to prove that the concave-upward surfaces merge with convex-upward rock floors, overlain by a thickening wedge of alluvium, at some distance from the range fronts.

In 1931, Johnson developed and greatly extended the concept of lateral planation as an alternate explanation of the extensive planed surfaces that had been attributed wholly or in large part to weathering, by num-

[63] *Op. cit.*, p. 449.

[64] Kirk Bryan: *Erosion and sedimentation in the Papago Country, Arizona*, U. S. Geol. Surv., Bull. 730 (1932) p. 19-90.

erous workers in the years that had intervened since the publication of Gilbert's views.[65] Shortly afterward, Blackwelder[66] ascribed to the "active sidewise cutting of desert torrents" the development of sloping planes of arid regions, having arrived at this view independently of Johnson.

In the 1931 article and in later papers, Johnson[67] pointed out the identity of Gilbert's local lateral corrasion surfaces with the extensive sloping rock planes brought to general attention by W J McGee[68] and others, and stressed specific inadequacies of the theory that these forms are produced by weathering and rill washing. He discussed the geomorphic evolution of a desert range in terms of three zones, in each of which the dominant action of streams differs from that in the other two:

> "(1) An inner zone, *the zone of degradation,* corresponding closely to the mountainous highland, in which vertical downcutting of streams reaches its maximum relative importance; (2) An intermediate zone, *the zone of lateral corrasion,* surrounding the mountain base, in which lateral cutting of streams attains its maximum relative importance. This is the zone of pediment formation. (3) An outer zone, *the zone of aggradation,* where upbuilding by deposition of alluvium has its maximum relative importance."[69]

As the two outer zones encroach on the wasting mountain mass, a planed rock floor, cut by swinging streams in the zone of lateral planation, tends to be buried by the advancing sheet of alluvium. If the rate of upbuilding of the detritus is rapid compared with the rate of retreat of the mountain front, successive layers will overlap the rock floor as it is produced, and the rock plane will slope more steeply than does the surface of the alluvium. But as the graded stretches of the streams are extended headward and as the supply of waste is decreased, the profiles of streams in the two inner zones are slowly lowered, in spite of continued upbuilding in the zone of aggradation. The exposed portion of the rock floor produced under these conditions is concave upward and carries only a thin veneer of alluvium, but the more deeply buried outer portion of the rock floor may be convex upward, thus corresponding to Lawson's suballuvial bench. Johnson calls particular attention to the importance of this constant regrading of the pediment slopes by swinging streams, and emphasizes the difficulty of producing and maintaining such slopes, adjusted to stream transportation, by any process other than stream erosion. He points out that dissection of pediments may be brought about by uplift or tilting, but that climatic change is equally competent to cause down-

[65] Douglas Johnson: *Planes of lateral corrasion,* Sci., n. s., vol. 73 (1931) p. 174-177.
[66] Eliot Blackwelder: *Desert plains,* Jour. Geol., vol. 39 (1931) p. 133-140.
[67] Douglas Johnson: *Rock fans of arid regions,* Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 389-416; *Rock planes of arid regions,* Geog. Rev., vol. 22 (1932) p. 656-665.
[68] W J McGee: *Sheetflood erosion,* Geol. Soc. Am., Bull., vol. 8 (1897) p. 87-112.
[69] Douglas Johnson: *Rock fans of arid regions,* Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 389.

cutting. Finally, he shows that landforms exhibiting the slopes of the alluvial fans, but cut in bedrock and mantled by a thin veneer of alluvium, should be formed by the lateral corrasion of swinging streams opposite canyon mouths and that the existence of such forms confirms the lateral corrasion theory of pediment origin.

Ross Field has recently reviewed the pediment problem, treating especially the form of the rock floor produced by lateral stream corrasion in a closed basin, under varying rates of mountain front recession and upbuilding of the central playa.[70] The importance of regrading is recognized, but Field seeks to make a sharp distinction in origin as well as in form, between the convex-upward rock floor, formed and deeply buried by alluvium during early stages of basin development, and the concave-upward, gravel-veneered plane, cut by regrading streams during later stages.

In a paper published after the first draft of this article had been completed, Rich attacks the theory of lateral planation, holding that pediments are extended by weathering back of the mountain front and are regraded by local wasting and sheetwash.[71] No general discussion of Rich's paper is attempted here, but it seems proper to point out that in each of the three theoretical cases analyzed by him the sharp break in slope between the mountain front and the pediment surface is controlled by structure or rock resistance. This critical angle has been recognized as a characteristic feature in all stages of erosion of arid ranges, in the general case where the rocks of the mountain front and the pediment are identical in structure and resistance, and its bearing on the two opposed theories of pediment formation has been discussed by Johnson,[72] Field,[73] and others. Whatever the merits of Rich's theory of pediment regrading by local processes in the Book Cliffs region, the evidence set forth in earlier sections of this paper (and hereinafter summarized) seems to prove that these processes play no essential part in the formation of gravel-capped surfaces in the Big Horn Basin.

LATERAL CORRASION PLANES OF MAJOR BIG HORN BASIN STREAMS

Origin.—Evidence bearing on the manner of origin of benches and terraces of the Big Horn Basin, as partial pediments, is reviewed briefly in the light of the foregoing discussion. Each surface is capped by rounded river gravels, composed of rock types that could not have been derived from the local bed rock. In each instance, the gravel is a sheet of mod-

[70] Ross Field: *Stream carved slopes and plains in desert mountains*, Am. Jour. Sci., 5th ser., vol. 29 (1935) p. 313-322.

[71] John L. Rich: *Origin and evolution of rock fans and pediments*, Geol. Soc. Am., Bull., vol. 46 (1935) p. 999-1024.

[72] Douglas Johnson: *Rock fans of arid regions*, Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 405-408.

[73] Ross Field: *op. cit.*, p. 317.

erate thickness, which exhibits original structures characteristic of fluvial deposits and invariably rests on a smooth rock surface that truncates structures of markedly different degrees of resistance. The bed rock underlying the gravel is characteristically fresh; no soil layers or weathered profiles have been found beneath the gravel cappings. The gravel-mantled surfaces are preserved at many levels, and each is a member of a complete sequence of analogous forms, decreasing in height and breadth to a floodplain now in process of formation by a major stream. Low scarps, which interrupt the graveled surfaces, are commonly associated with cut steps in the rock floor on which the gravel rests, and, in plan view, these scarps are identical with those produced by sweeping and swinging meanders on present floodplains. Distribution of distinctive rock types in the gravel sheets on high divide benches makes it possible to identify each bench with the present major stream by which it was formed at a higher level. Each gravel sheet, from those of the present floodplains to those of the highest benches, is overlain by a thin covering of floodplain silt, and the higher, and older, sheets are made up, in part, of completely decomposed pebbles, which still retain their exterior form. These relations, and others already mentioned, demonstrate that the gravel sheets are now in essentially the same state as when deposited by major streams, and that the effect of creep, sheet wash, and other local processes in lowering or otherwise modifying the flat tops of the benches is negligible. These local processes do tend to destroy the gravel-mantled planes, by removing the weaker rocks on which the gravels rest, producing steep escarpments and rock-ribbed badlands wherever minor differences in rock resistance are present.

Relations of gravel thickness and the form of the rock floors.—The lateral planation concept has been developed largely in the Southwest, and illustrations have been drawn almost exclusively from that region of relatively recent differential movement, where streams issuing from arid mountains may lose themselves on broad detrital plains, or end in playa flats of tectonic basins, or enter valley floors that carry integrated basin drainage to the sea. In the first and second examples, upbuilding of stream courses beyond the range fronts usually accompanies reduction of the mountain mass, so that, as the pediment is extended, its distal end is buried under a thickening alluvial mantle. As already explained, such accidents as climatic change, uplift, or basin capture may cause degradation locally or throughout the drainage system, or, in the absence of such interruptions, lowering of stream profiles in the normal course of the cycle may cause downcutting (regrading) for some distance outward from the mountain front, in spite of continued upbuilding in the lower reaches of the streams; in either instance, a concave-upward rock floor,

carrying only a thin veneer of alluvium, is produced at the inner margin of the earlier-formed and more deeply covered plane. Numerous published cross-sections of southwestern pediments show that, as might be expected, a combination of the convex-upward rock plane, overlain by a thickening alluvial wedge, and the concave-upward gravel-veneered surface is the typical form.

Conditions are wholly different in the Big Horn Basin, where the activity of streams issuing from the mountain fronts has been controlled throughout the present cycle by the stable or slowly lowering base level provided by through-flowing master streams, fully competent to transport available waste far beyond the limits of the region. The direct result of this essential difference is that the convex-upward floor and its associated alluvial wedge are wholly absent in the basin, where the surfaces cut by stable or slowly degrading streams are, like the profiles of the streams, concave upward and mantled by alluvium equal in thickness to the depth of effective flood scour.[74]

It is evident that this contrast is one of stage rather than of place or kind. Certain pediment forms figured by Johnson and others described by Bryan may be essentially similar to those of the Big Horn region, in that the surfaces were produced by degrading streams and truncate structures of older alluvium. The convex-upward rock floor may have been formed in the Big Horn Basin during later stages of the period of widespread middle and late Tertiary aggradation, when surfaces of low relief, still mantled by stream gravels, were developed on high shoulders of the Big Horn Mountains. The surfaces discussed in this article were produced during pauses in a cycle of deep regional degradation, totalling approximately 4,000 feet, which followed the episode of basin filling.

Longitudinal extent.—The planed rock floors of southwestern regions are extended largely by headward retreat of the mountain fronts and often bevel weak and resistant rocks alike, but the gravel-capped surfaces now preserved in the Big Horn Basin were opened outward, by selective lateral corrasion in weak rocks of the basin; major streams enter and leave the basin through relatively narrow canyons cut in resistant rocks of the ranges. Here, as in the matter of the form of the rock floor, it is important to note that the contrast is one of stage rather than of kind, for, other things being equal, the extent of a lateral corrasion surface is inversely proportionate to the resistance of the country rock and directly proportionate to the duration of lateral cutting at one level. The several pauses in downcutting in the Big Horn Basin have lasted long

[74] The same thought, expressed in almost identical terms, appears in a recent paper by Kirk Bryan and Franklin T. McCann [*Successive pediments and terraces of the upper Rio Puerco in New Mexico*, Jour. Geol., vol. 44 (1936) p. 199].

enough for the streams to cut extensive planes in weak rocks, but each of the stages of dominant lateral planation recorded by the benches and terraces was interrupted by downcutting, before the valley floors were widened appreciably in resistant rocks. But, as already shown, it is probable that during earlier stages the pediments of the basin did encroach on the surrounding mountain masses, and it is theoretically certain that, if downcutting were to cease for a sufficiently long time, complete planation of the basin floor might be followed by slow extension of the cut plane toward the centers of the ranges, as a result of continued lateral cutting by streams.

Lateral extent.—Southwestern pediments are typically broad surfaces of complete planation at, or little above, present stream levels. The major streams of the Big Horn Basin occupy relatively narrow valley floors flanked by stream-cut rock terraces; higher benches on interstream divides are certainly remnants of once more-extensive lateral corrasion planes, but positive determination of their former breadth is difficult or impossible. Distribution of Rock Creek gravels at several levels more or less remote from the present stream proves that the minimum breadth of lateral "migration" of Rock Creek during several post-Mesa stages is approximately 7 miles at a point 13 miles distant from the mountain front, or equivalent to an arc of 35 degrees. The minimum breadth of Greybull "migration" at, or below, the Tatman level, as determined by distribution of distinctive Greybull gravels, is 14 miles at a point 44 miles from the mountain front, or equivalent to an arc of about 20 degrees. It should be emphasized that these are minimum figures, for the gravel-capped benches and terraces on which they are based have been greatly reduced by erosion. Indirect evidence bearing on the original extent and on the form of these or similar surfaces at higher levels is provided by the patterns of many of the small streams of the basin.

Willow and Red Lodge creeks, minor streams that drain the lowland area west of the Rock Creek valley, rise near Rock Creek but diverge from it in courses that radiate from its canyon mouth in the Beartooth Mountain front (Pl. 10). Elbow Creek and the gravel-mantled Mesa Bench remnants, east of the present valley, diverge from it in the same manner. A strikingly similar sequence of radial elements is in the Greybull area, where Grass, Gooseberry, Dry Cottonwood, Elk and Dry creeks and numerous bench and terrace remnants north and south of the present Greybull Valley diverge spoke-like from the point where the Greybull issues from the Absaroka Mountains (Pl. 11). In both instances, the radial streams cut directly across rock structures, and it appears that the only rational explanation of their courses is that they were determined by the radial slopes of broad valley floors formed by the main

streams at some level, or levels, well above the present surface. The possibility that the old valley floors were aggradational forms, built during the basin-filling episode, cannot be eliminated by direct evidence, but the presence of Tatman Mountain, the Mesa Bench, and other remnants of stream-carved, gravel-mantled surfaces, at random positions between the radial streams, affords strong support to the theory that the valley floors from which the minor streams were let down were cut surfaces, formed by lateral corrasion by the main streams, during pauses in the degradation of the basin. If such were the case, the indicated minimum breadth of Rock Creek "migration" is increased from the 35 degrees of arc proven by gravel distribution to 55 degrees and that of Greybull "migration" from 20 to 60 degrees.

Theoretical considerations, independent of the diverging pattern of the supposed radial consequents, indicate that a valley floor of such great breadth, cut by a stream issuing from a narrow canyon mouth, could not be flat in cross profile. Just as, on a broad alluvial surface formed opposite a mountain gorge by an aggrading stream, each part of the fill surface at the same distance from the point of discharge must be at the same, or nearly the same, elevation, so, on such a cut surface, all points at an equal radial distance from the canyon mouth must be at the same level; contour lines crossing either surface must bow outward as arcs of concentric circles described from the canyon mouth. The radial slope of the surface is, in both instances, controlled directly by the declivity of the swinging stream and is concave upward; its convexity in cross profile is, in both instances, a function of the declivity of the stream and the breadth of the surface. If the form is depositional, produced by an aggrading stream, it is an alluvial fan or cone; Johnson has shown that if it is erosional, produced by a stable or slowly degrading stream, it is a rock fan.[75]

Whether the floor of the Big Horn Basin has been completely planed across by swinging streams, at one or more times during the period of degradation, is not certainly known. The great breadth of the fan forms defined by radial stream courses opposite Rock Creek and Greybull canyon mouths indicates that, if it had been so planed, the basin floor must have been essentially a series of coalescing rock fans, each apexing at the canyon mouth of a major stream and all sloping to the broad valley floor of the master Big Horn River.

Stream diversion.—Relations between adjacent major streams on such broadly opened surfaces as those described have received scant attention in the literature on geomorphology, but this phase of the pediment prob-

[75] Douglas Johnson: *Rock fans of arid regions*, Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 389-416.

lem is of especial importance in the Big Horn Basin. The large streams issuing from the mountains at the western margin of the basin differ markedly in declivity. For this reason, the pediment segments or rock fans produced by several streams during any one pause in downcutting should be at the same, or nearly the same, level near the master streams but should diverge in level upvalley, or toward the mountain front. It is evident that if, during any stage of broad planation, the last fringing divide between two adjacent streams had been completely destroyed by the lateral cutting of the stream of lower declivity, a scarp would be produced, its height depending on the difference in the declivities of the streams and the distance from their mouths. If the higher-gradient stream were then to swing laterally to the vicinity of the scarp, it would almost certainly spill into the lower valley of its neighbor, would regrade itself to the new local base level, and would thereafter, in the normal course of events, free itself with considerable difficulty from its vassal state.

It might be argued, therefore, that the fact that the high-gradient Greybull and the relatively low-gradient Shoshone, for instance, pursue individual courses from the Absaroka Mountain front to the Big Horn River, proves that the divide between the two streams has never, during the degradation of the basin, been completely cut away; during earlier stages of the study reported here, this consideration made the writer reluctant to admit the possibility of complete planation, in spite of the presence of gravel-mantled, stream-planed surfaces on the divides between each of the major streams of the basin. A complete answer to the seeming difficulty is provided by a second type of stream diversion, illustrated by the recent capture of the Greybull River.

It has been demonstrated that the Greybull was captured because its graded declivity, adjusted to the transportation of coarse rock waste from the Absaroka Mountains, was higher than that of a much smaller neighboring stream, which carried only fine silts and sands derived from weak basin-filling strata. Similar conditions favoring the capture of large streams issuing from the mountains by small basin streams have existed throughout the degradation of the basin, and it is apparent that such diversions[76] provide a ready mechanism by which the effect of captures of the first type (namely, the coalescence of pediment-forming streams by intercision) may be undone.

Capture by intercision must be common on all broadly planed pediment surfaces. In addition, captures of the Greybull type may be expected wherever the rocks of the pediment differ markedly in resistance from those of the mountain mass. Captures of both types have probably occurred repeatedly during the degradation of the Big Horn Basin,

[76] Similar captures have been described in the Book Cliffs region, in Utah [John L. Rich: *Origin and evolution of rock fans and pediments*, Geol. Soc. Am., Bull., vol. 46 (1935) p. 999-1024].

and it appears that the somewhat involved sequence of drainage changes described in earlier sections must be considered normal or merely usual, rather than abnormal.

The term "lateral migration" rather than "lateral swinging" has been used in discussion of the significance of the distribution of gravels and radial stream courses, because it is evident that streams may be shifted laterally on fan or pediment surfaces, by intercision and by capture of the Greybull type, as well as by lateral swinging.

LATERAL CORRASION PLANES OF MINOR STREAMS

The field studies on which this report is based were confined largely to the gravel-mantled benches and terraces formed by major streams issuing from the western marginal ranges of the Big Horn Basin; surfaces produced by minor streams that rise in the basin proper or descend into the basin from the surrounding mountain fronts were examined only incidentally in the course of the larger work. They are described briefly, because they are, in some instances, identical in appearance with the master-stream surfaces and might be confused with them, and because they reproduce, on a smaller scale and in almost infinite variety, the forms related to the activity of the master streams. The minor-stream planes vary in declivity from 80 to 600 feet per mile; they vary, also, in breadth and in the perfection of the development of rock fans at their rear, or "mountainward", margins. They are, like the main-stream lateral corrasion planes, concave upward in longitudinal profile and are overlain by a uniformly thin mantle of alluvium,[77] this being the direct result of the fact that the cutting of the minor streams is controlled by graded or slowly degrading master-streams, to which they contribute.

These forms are most typically developed around the retreating margins of residual highlands on the Clark Fork-Shoshone divide, in the west-central part of the basin. The Eaglenest Creek Flat (Fig. 11) slopes smoothly from the southeastern base of Heart Mountain to the level of the Powell Terrace, in the Shoshone Valley, and was evidently formed during the Powell stage of regional erosion. The surface truncates structures of Eocene sandstones and shales and is mantled by alluvium containing abundant limestone fragments, deposited by Eaglenest Creek, Alkali Creek, and other minor streams heading against the Heart Mountain overthrust blocks. Some small streams have trenched the surface, and their inner valley sides provide excellent sections, showing the planed rock floor overlain by from 10 to 15 feet of stream detri-

[77] Detrital material washed down from cut scarps over the minor stream gravel sheets is left out of account here, as in the earlier discussion of the terraces of the Shoshone and other large streams of the basin. Slope wash may form a smooth slope from a weathered meander scarp to the gravel-mantled plane at its base, obscuring the angular break between the two unlike forms.

tus. Repeated measurements of gravel thickness justify the statement that the sub-alluvial floor is essentially parallel with the surface of the

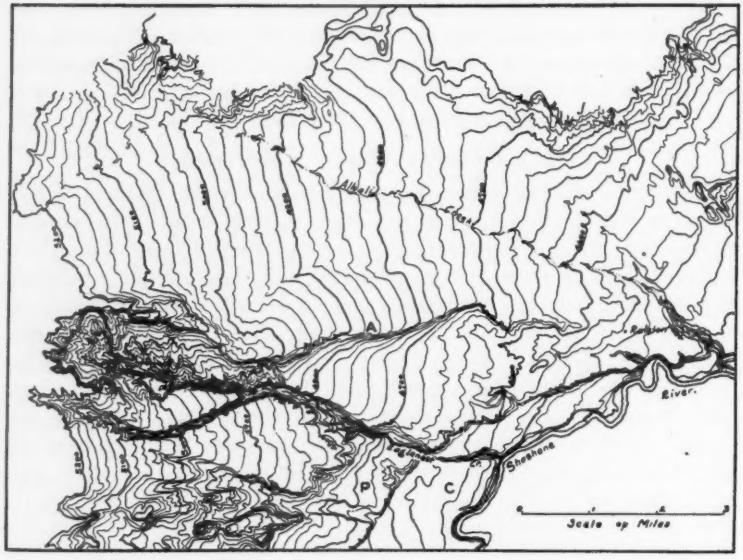

FIGURE 11.—*Topographic map of the Eaglenest Creek "Flats"*

Cody (C) and Powell (P) surfaces are stream-cut rock terraces, formed by the Shoshone River and veneered by a thin mantle of typical Shoshone River gravel. The terrace treads are deeply covered by slope wash, but the main-stream gravel sheets and the planed rock floors slope downvalley (northeastward). The broad flats traversed by Alkali and Eaglenest creeks slope toward the main-stream valley. A smoothly planed bed-rock surface, truncating structures and overlain by a uniformly thin mantle of stream-borne detritus containing much limestone, is exposed for several miles in the inner valley sides of Eaglenest Creek and in the scarp running through A. The surface trenched by Eaglenest Creek is a compound, or "two story," rock fan. (Map, United States Reclamation Survey, 1903.)

alluvium, over the greater part of the "Flat"; the distinct fan incised by Eaglenest Creek is cut in rock, not built of alluvium.

Chapman and Kimball benches slope northward from the northern bases of Heart Mountain and Polecat Bench to the edge of low meander scarps overlooking the present channel of the Clark Fork River, and their riverward margins, viewed from the opposite bank, might well be mistaken for gravel-mantled main-stream terraces. But the gravels are composed almost wholly of limestone and agglomerate pebbles derived from Heart Mountain and the Polecat gravel sheet; whereas the Clark Fork gravels are rich in distinctive granites and gneisses from the Beartooth Mountains. The sub-alluvial surface exposed in tributary trenches

slopes toward the main stream, not down-valley. It appears, therefore, that the perfect planation of the rock floor on which the gravel rests was accomplished by the side streams by which the gravel was deposited. The minor benches have not been surveyed and are not indicated on maps that accompany this paper.

The manner in which these broad surfaces are extended headward can be readily understood, even in dry weather when the minor streams are not operating. From a vantage point on any of the retreating scarps, the much lower surface sloping outward from the base is seen to be covered by a network of dry stream channels, revealed by distribution of vegetation and reworked bench gravel. The scarp face is dissected by the ramifying headwater gullies of many streams, and its base is embayed by extensions of the broad sloping plane up the valley of each minor stream, or, conversely, the plane is formed by the coalescence of a multitude of small valley floors. The manner in which the valley floors are widened, and the tapering interstream spurs are cut away, is indicated by marked steepening or undercutting of the valley sides wherever a stream has recently swung against them. The extension of the plane and the retreat of the scarp at constant declivity appears to be the result of two unlike processes; first, constant weathering and wasting of the scarp face, which tends to reduce its declivity, and, second, intermittent stream cutting at the base, which tends to steepen it.

Limits of space forbid more than brief mention of a few additional examples of surfaces formed by minor streams in the basin area. Hewett has described gravel-mantled planes in the Meeteetse district and has shown, by curving contour lines, the fanlike form of the surfaces.[78] He fails, in his published account, clearly to distinguish surfaces formed by the Greybull River from surfaces cut by Meeteetse Creek and other minor streams that join the main stream. In perfection of beveling and in uniformity in the thickness of the capping gravels, the minor-stream corrasion planes rival those produced by the main Greybull River, but the surfaces cut by the side streams slope far more steeply than those of the Greybull, and they slope toward the main valley.

Grove Creek Bench, Line Creek Bench, and a number of other planed rock surfaces slope, with varying declivity, from the Beartooth Mountain front to the segment of the Clark Fork River just north of its canyon mouth. The surfaces are overlain by a mantle of alluvium, rich in distinctive porphyries from the adjacent part of the Beartooth range, and were clearly formed by the cutting of the local streams. As the projections of the several concave-upward surfaces meet in a common plane

[78] D. F. Hewett: *Geology and oil and coal resources of the Oregon Basin, Meeteetse and Grass Creek Basin quadrangles, Wyoming*, U. S. Geol. Surv., Prof. Pap. 145 (1926) p. 6.

200 to 300 feet above the level of the Clark Fork River, it appears that the benches are of the same age; marked differences in the level at which they approach the mountain front are due to differences in the declivities of the minor streams by which the benches were cut and to differences in their length.

The broad surface of Wagonhound Bench, opposite the mouth of Cottonwood Creek, in the southwestern part of the basin, would, from the evidence of the topographic map alone (Grass Creek Basin quadrangle), be interpreted as a segment of a regraded and dissected alluvial fan, but study of the lateral bench margins proves that the alluvium is a uniform layer, resting on a surface that bevels bed-rock structures. A minor Greybull type capture, by which Cottonwood Creek has been transferred from the northern segment of Wagonhound Bench to its present valley, is of particular interest.

PROBLEM OF LAND MOVEMENTS DURING PERIOD OF DEGRADATION

General statement.—It has been suggested by several writers already cited that the gravel-mantled benches that slope outward from many of the ranges of western Wyoming are peneplanes. If this be true, the present steep plainsward slopes of the bench surfaces prove that the ranges have been affected in recent time by notable elevatory movements accompanied by warping of the plains border, for peneplanes are formed with low regional slopes. But it has been demonstrated that the benches of the Big Horn Basin were produced by the lateral cutting of streams of high declivity and that their slopes differ only slightly from the slopes of valley floors now in process of formation. The gravel-capped surfaces are remnants of stream-cut valley flats, not peneplanes; Douglas Johnson has shown that it is wholly unnecessary to postulate marked differential movements to account for the slopes of such surfaces.[79]

As stream-planed surfaces, the terraces and benches of the basin present two related problems, the first being the reason for the downcutting indicated by successively lower levels of planation, and the second the reason for the divergence of bench and terrace profiles from the profiles of present valley floors.

Problem of regional uplift.—One outstanding fact, clearly recognized by earlier workers and confirmed by the present study, is that degradation of the streams of the Big Horn Basin has been pulsational, at least four stages of dominant planation, separated by periods of more rapid downcutting, being recognizable. Depths of entrenchment between planation levels are commonly held to record changes in the relative height of the Big Horn region above the controlling base level of the sea, by nearly

[79] Douglas Johnson: *Rock fans of arid regions*, Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 410.

equal amounts and during separate elevatory movements between distinct pauses in uplift.

This simple and direct relation between stream trenching and regional uplift in the Big Horn Basin is valid, provided: first, that the graded declivity of the Big Horn-Yellowstone-Missouri (Mississippi?) river system has been constant during the cutting of the several surfaces, and, second, that the mouth of the master stream has been more or less fixed in geographic position. Pulsational downcutting of Big Horn Basin streams may be held to record per saltum changes in the actual or absolute level of the land, only if it can be further demonstrated that the level of the sea has been constant during the period of entrenchment. Each of these assumptions is either of doubtful validity or is contrary to known fact. The westward divergence of profiles of high-level terraces from the present graded profiles of the master streams of the Northern Plains is open to two interpretations, one of which is that the graded declivities of the streams have decreased. Alden and others have shown that the course of the Missouri River has been repeatedly blocked and diverted by Pleistocene ice and that the drainage of the Big Horn Basin region as a whole has been transferred from Hudson Bay to the Gulf of Mexico.[80] The level of the sea has fluctuated in late Tertiary and Pleistocene time.

In view of these facts, the four, or more, distinct pauses in the downcutting of Big Horn Basin streams cannot, in the opinion of the writer, be considered compelling evidence of an equal number of pauses in regional uplift. So many factors other than land movements may, by altering relations of load and volume, cause pauses or quickenings in the downcutting of graded streams in the center of a broad continental area that positive assignment of such changes in stream activity to pulsational movements of the land alone seems in the highest degree unsafe.

The question as to the validity of pulsational stream downcutting as a proof of pulsational uplift leads to the broader question as to whether or not the 3000 to 4000 feet of downcutting in the present cycle in the Big Horn Basin must be ascribed to, and accepted as proof of, a regional uplift of equal or greater amount since degradation began. The theory that the Rocky Mountains and the western Plains have been notably uplifted in late Pliocene or Pleistocene time is frequently expressed in geological papers and appears to be generally accepted. The theory is based on the recognition of supposed peneplane surfaces of mid-Tertiary age beveling range crests, and the difference in elevation between the postulated level of peneplane formation and the present level of the peneplane surface, or surfaces, is considered to record the amount of uplift.

[80] W. C. Alden: *Physiography and glacial geology of eastern Montana and adjacent areas*, U. S. Geol. Surv., Prof. Pap. 174 (1932) 133 pages.

Blackwelder, in an article dealing with the history of the Colorado River, states:

> "There is ample evidence, noted by many writers on the region, that uplift amounting to thousands of feet—perhaps as much as six thousand feet in the central Rocky Mountain region—did occur sometime after the Pliocene period. It was this broad warping that induced the cutting of countless gorges and canyons throughout the province and caused the ranges themselves to be etched into greater relief."[81]

Douglas Johnson, however, has suggested that the smoothly sloping erosion surface at 7000 to 9000 feet in the Colorado Rockies may be a pediment rather than a peneplane, and that, as such, it might have been formed at a high level.[82] Directly opposed to theories that involve reduction of the Rockies to a low-lying peneplane in mid-Tertiary time, and in harmony with Johnson's hypothesis, are the several lines of evidence discussed in the first section of this paper—namely, the essential similarity of mid-Tertiary and present semi-arid climates of the Great Plains, and the necessity of high stream declivities to explain transportation of alluvial materials of the mid-Tertiary High Plains mantle eastward from the Mountains.

The latter point may be emphasized by calling attention to the prevailing coarseness of the detritus shed from the Northern Rockies during various stages of the Tertiary. Conglomerates interbedded with fossiliferous Oligocene sands in the Cypress Hills are certainly Oligocene, whatever the age of the Plateau surface. They consist of pebbles up to 12 inches in diameter, and they are preserved at points from 150 to 200 miles east of the mountain front. Lambe,[83] Alden,[84] and others have pointed out that these gravels must have been transported by turbulent streams of high declivity, and that the headwater basins from which they were derived must have stood at a high level. Supposedly late Miocene or Pliocene Flaxville gravels, and the gravels of lower terraces of Pliocene and Pleistocene age, are widely distributed over the Montana Plains. They differ from the Cypress gravels (as defined in this paper) in that they were deposited as thin sheets on stream-cut rock terraces during the current period of degradation; whereas the thick Cypress sequence was probably formed by aggrading streams during the preceding period of upbuilding; but they are uniformly coarse. It appears, therefore, that the conclusions of Lambe and of Alden, as to the high declivities of Oligocene streams, must be extended to include Miocene, Pliocene, and Pleistocene times. The streams of the Northern Plains, like those of the

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[81] Eliot Blackwelder: *The origin of the Colorado River*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 561.
[82] Douglas Johnson: *Rock fans of arid regions*, Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 410.
[83] L. M. Lambe: *The vertebrata of the Oligocene of the Cypress Hills, Saskatchewan*, Canada Geol. Surv., Contr. Pal., vol. 3, pt. 4 (1908) 65 pages.
[84] W. C. Alden: *op. cit.*, p. 6.

Central Plains area discussed by Willard Johnson, had high declivities throughout middle and late Tertiary time, because they were heavily loaded, and the load must have been derived from a Rocky Mountain range that was consistently high.

This general conclusion will be elaborated further in later papers; the immediate question here under consideration is whether downcutting in the present cycle in the Big Horn Basin is due to regional uplift, or whether it might be otherwise explained. The foregoing discussion, which tends to show that the Rocky Mountains and the western Plains border stood at a high level before degradation began, establishes the question as a real, not an academic, one.

A major factor that bears directly on the problem of the cause of downcutting is the possible effect of late Tertiary and Pleistocene climatic changes on the activity of streams of the Plains. Brooks [85] and others have suggested that the presence of extensive ice caps over continental areas might be expected to cause marked changes in atmospheric circulation over the whole surface of the earth. Gilbert,[86] Johnson,[87] and others have shown that periods of glaciation correspond with pluvial stages in normally arid or semi-arid regions well beyond the ice borders, and Simpson [88] has advanced a thought-provoking hypothesis, according to which, pluvial stages in ice-free regions are differently spaced, but closely related, to recurrent glacial stages. Even if possible increases in precipitation are left out of account, it seems probable that general Pleistocene refrigeration of the lands, especially near continental ice caps, might, by decreasing evaporation, tend to cause an increase in runoff during glacial stages. It seems reasonable to believe that the general effect of such changes would be to cause streams flowing eastward from the Rocky Mountains during late Pliocene and Pleistocene time to regrade their courses across the Plains at a lower declivity, as a result of a relative increase in volume. Downcutting resulting from such changes might be expected to be more marked in the Northern than in the Southern Plains region, as derangements of atmospheric circulation and temperature changes would have been especially effective near the ice front; there is abundant evidence that the Northern Plains are more deeply eroded than are the Central or the Southern Plains, and it has been shown that the apparent conflict in the age of topographic forms in northern and southern regions may be accounted for by a southward-working wave of rejuve-

[85] C. E. P. Brooks: *Climate through the ages*, London (1926).

[86] G. K. Gilbert: *Lake Bonneville*, U. S. Geol. Surv., Mon. 1 (1890) p. 265-316.

[87] Willard Johnson: *The High Plains and their utilization*, U. S. Geol. Surv., 21 Ann. Rept., pt. 4 (1901) p. 655.

[88] G. C. Simpson: *World climate during the Quaternary period*, Roy. Meteorol. Soc., Quart. Jour., vol. 60 (1934) p. 1934; reviewed by Richard Foster Flint: Am. Jour. Sci., 5th ser., vol. 29 (1935) p. 381-385.

nation, for which there is independent evidence. Downcutting resulting from climatic changes related to glacial stages should be interrupted by pauses or reversals during interglacial times; it has been shown that downcutting in the Big Horn Basin has been pulsational.

The conclusion to be drawn from this reasoning is not that the Big Horn region has not been affected by a number of per saltum uplifts, totaling 4,000 feet, or more, during the present cycle of erosion, but, rather, that the physiographic forms of the Basin do not, in themselves, constitute proof that such pulsational uplift has occurred. Regional halting uplift of the Rocky Mountains and the Great Plains, greater at the north than at the south, is certainly competent to explain observed variations in stream activity, but the relations may be equally well explained, in part, at least, as the results of climatic change in late Tertiary and Pleistocene times. Until it is demonstrated that the graded declivity of the Tertiary master stream that drained the area before downcutting began could not have been 4 feet per mile steeper, in its 1,000-mile course to the sea, than the graded declivity of the present master stream, the possibility that the basin region has not been notably uplifted in the present cycle must be entertained. The single interpretation of stream trenching that is fully justified by observed relations in the local Big Horn Basin area is that the trenching is due, in part, to downcutting of the Yellowstone River immediately to the north.

Problem of differential movement.—The declivity of a graded stream is controlled solely by the relation of the volume of the stream to the amount and character of its load. Simple lowering of a local base level should cause a graded stream to cut downward, but, other factors remaining the same, the new graded profile should not differ in declivity from the old, except insofar as quickened downcutting alters the balance of load and volume. Differences in the declivities of terraces at different levels in the same valley segment must, therefore, indicate either warping of the land during the period of terrace formation or changes in relations of load and volume, which have caused the stream to alter its graded declivity in adjusting itself to the new set of controlling conditions. The problem of distinguishing between the effect of differential movement and that of changes in stream declivity, in the interpretation of the longitudinal profiles of Big Horn Basin terraces, presents formidable difficulties.

The prevalence of upvalley divergence throughout the Big Horn Basin and in adjacent areas suggests that this feature may be the result of some regional change, such as a variation in climate, which has, by altering relations of load and volume, caused the streams to regrade their courses to a lower declivity since the upper terraces were formed. Closely related to climatic change, but differing from it in the manner in which

it affects stream declivities, is the advent of local glaciation. No extended discussion of the possible effects of the accumulation of ice in the headwaters of streams adjusted to ice-free conditions is attempted here; but it does seem probable that the earlier delicate balance of load and volume would be changed, and that the effect might be an increasing or a decreasing of the preglacial slope, or both increasing and decreasing during different phases of ice activity, or increasing or decreasing after the ice had retreated from valleys choked with morainal debris.

An exception to the general rule of upvalley terrace divergence occurs in the Rock Creek valley, where the Mesa Bench and the Roberts Terrace slope less steeply than does the lower Red Lodge Terrace. Changes in the declivity of Rock Creek, of the kind necessary to account for exceptional variations in terrace slope in the Rock Creek valley (without warping), may be due to local causes, such as a decided loss of tributary volume, resulting from westward retreat of the Clark Fork divide and recent diversion of tributaries of the West Branch of Rock Creek by Red Lodge Creek, or the changes may be due to the general regional changes already noted.

Variations in the slopes of terraces at different levels in Big Horn Basin stream valleys may be equally well explained by warping of the plains border adjacent to the Beartooth-Absaroka mountain front. The fact that the general upvalley divergence of high-level terraces is not quantitatively proportionate to the declivities of the several streams seems to favor variable warping, for climatic changes, which alter relations of load and discharge of all the streams of an area, should cause proportionate changes in their graded declivities. The further fact that the Cody and the Powell terraces in the Shoshone Valley and the Red Lodge Terrace in the Rock Creek valley begin to diverge from the profiles of the present stream at points more or less equidistant from the mountain front, rather than at the mouths of the streams, supports the warping theory, for changes in graded declivity would presumably extend throughout the lower reaches of the streams.

Faulted alluvial fans and moraines near range fronts in the Rocky Mountain region beyond the borders of the Big Horn Basin provide positive evidence of upward movements of the ranges relative to the surrounding plains in recent time. Duncan Johnson has described relations that suggest recent movement along the west marginal faults of Rattlesnake Mountain, west of Cody, and minor earthquakes have been reported on the western slopes of the Big Horn Range.[89] These general considerations indicate that warping of the floor of the intermontane Big Horn Basin

[89] C. Duncan Johnson: *Geology of the mountain uplift transected by the Shoshone Canyon, Wyoming*, Jour. Geol., vol. 42 (1934) p. 833.

is not inherently improbable. Any attempt to fix the amount of the warping, in a quantitative sense, by mountainward divergence of high-level stream terraces must be based, however, on the assumption that the graded declivities of Big Horn Basin streams have been constant throughout the period of terrace formation, and, in view of known variations in conditions controlling relations of stream load and volume in the recent past, this assumption is untenable. Probably, both warping of small amount and changes in graded stream declivity have occurred in the basin region; the evidence in hand does not, in the writer's opinion, justify assignment of any particular case of variation in terrace slope to either cause alone, nor any statement as to the quantitative importance of either.

SUMMARY

The Big Horn Basin is one of a series of intermontane basins that characterize the Rocky Mountain front in western Wyoming. The floor stands approximately 4,000 feet above sea level, and the surrounding ranges rise to heights of from 8,000 to 13,000 feet. The climate is arid to semi-arid.

Paleontologic evidence indicates that the basin floor stood at a relatively low level during the period of Laramide differential movement in which the structural depression was formed. Prevailing Paleocene and early Eocene humidity suggests that the early Tertiary relief of the Big Horn region was less than the present relief. Marked changes in the direction of increasing aridity on the Great Plains and in intermontane basins in later Eocene and Oligocene time, interpreted here as the direct result of broad regional uplift of the Rocky Mountains, caused widespread aggradation by eastward-flowing streams, which culminated in mid-Tertiary time in the production of the vast alluvial plain now preserved in the High Plains surface. The Rocky Mountains presumably were not reduced to a low level, relative to the sea, during the mid-Tertiary, for the aridity of the Plains area to the east, doubtless related to the present height of the mountains, was quite as marked during mid-Tertiary time as at present. The coarse detritus spread eastward across the Plains during Miocene and Pliocene times could not have been derived from a low-lying peneplane surface.

The present cycle of degradation was apparently initiated during the Pliocene in the Central Plains area, possibly at an earlier date on the Montana Plains and in Canada, and at an indeterminate time, possibly late Miocene or early Pliocene, in the Big Horn Basin.

Residual landforms marking pauses in the re-excavation of the basin during the present cycle of downcutting are high, gravel-capped benches on interstream divides and broad terraces in stream valleys. They were

formed by lateral corrasion of graded streams issuing from the encircling mountain fronts and are to be considered as remnants of broad valley floor flats, or partial pediments, rather than peneplane remnants.

Downcutting of 3,000 to 4,000 feet in the present cycle does not necessarily prove regional uplift of equal or greater amount since degradation began, for a decrease in the graded declivities of master streams flowing to the sea would produce a similar effect. Nor can pauses in stream downcutting, recorded by benches and terraces, be held to record pauses in regional movements of the land, for many factors other than land movements may cause pauses or quickenings in stream downcutting in the center of a broad continental area.

Differences in the declivities of bench and terrace profiles from those of profiles of present valley floors may be due to warping after the higher surfaces were formed or to changes in the graded declivities of the streams. Assignment of any particular case of terrace divergence to either cause, and exclusion of the alternate possible cause, is not justified by the evidence in hand; but it may be stated with certainty that, if warping of the basin floor has occurred during the period of bench and terrace formation, it has been slight in amount and has not necessarily been pulsational.

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